QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Editor: G. D. ROBINSON

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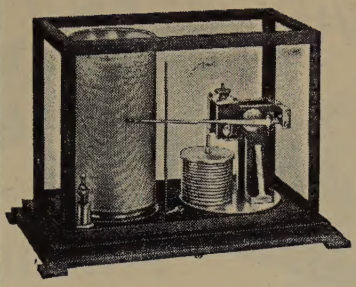

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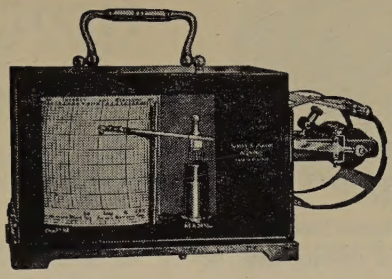

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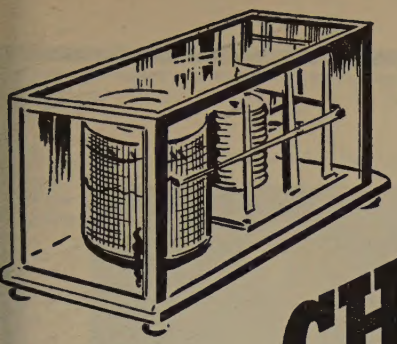

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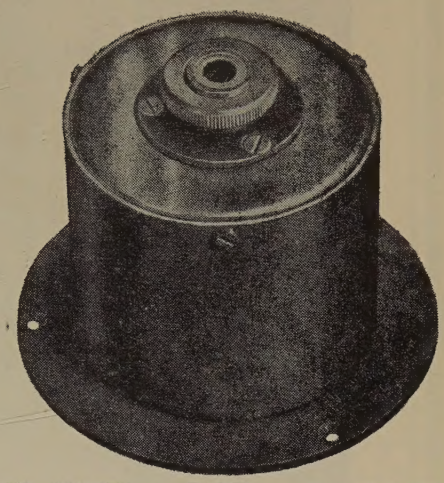

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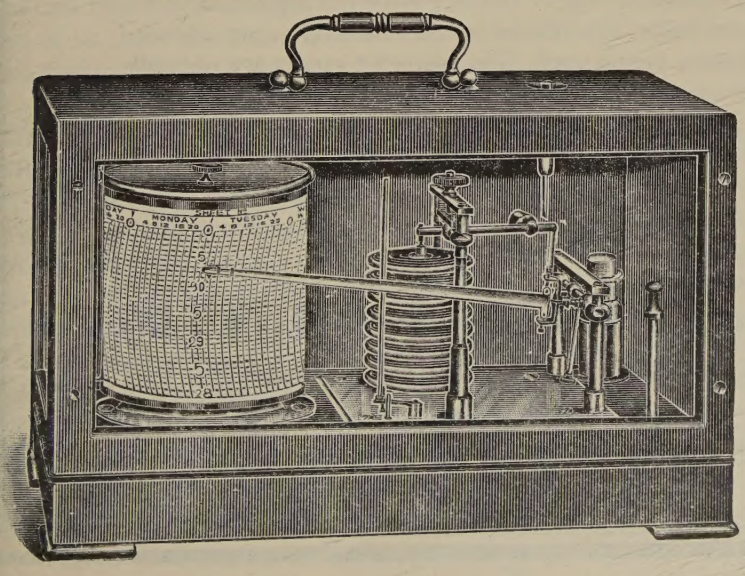

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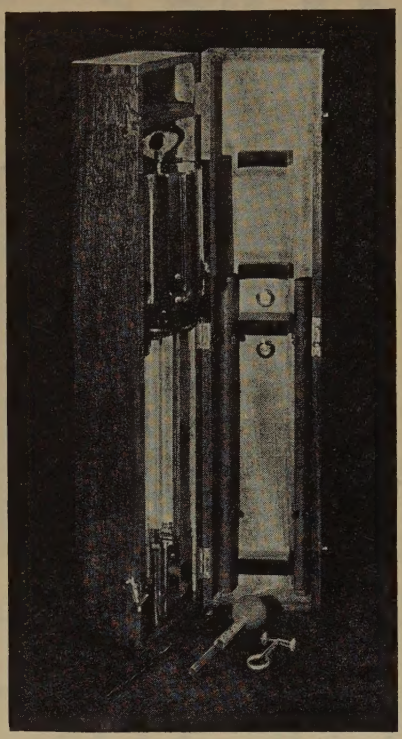

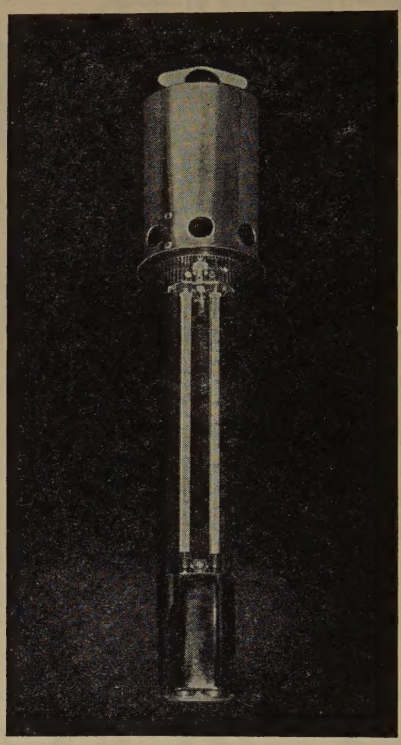

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An outlook on meteorology and the Society

By R. C. SUTCLIFFE, O.B.E., F.R.S.

Presidential Address delivered before the Royal Meteorological Society on 17 April 1957

It will be clear from the title of my Address that I intend to take this opportunity to offer some general observations on the state of our Society and of our science. I make no apology for an Address of this nature since I believe that the occasion is timely for a little stocktaking. It is almost impossible to pick up a newspaper or periodical at the present time without being reminded that our civilization has in recent years entered a new phase, the scientific technological age, with scientific research now playing a leading role. Since this Society has no small measure of responsibility for the performance of a branch of natural science of some consequence, it is right that we should spare a little time to consider our position.

This is not the place to expand upon the characteristics of the new technological age. Some will deny any recent change, asserting that the steady improvement in material living standards over at least a hundred years has been possible only as the result of scientific and technological progress. In this assertion they are undoubtedly correct but there is nevertheless an essential change: we have recently become self-conscious about these things. Not only the scientists and the technologists themselves but also the economists, industrialists, educationalists, politicians and above all, Governments, have realized that for the improvement of material living standards there must be an efficient use of our scientific resources. It may be that this improvement is a worthless objective but public opinion is convinced it is worthwhile and such is the spirit of the age.

The great expansion in scientific research has come in the post-war years and meteorology has received its share; compared with the pre-war period, the increase in research effort must have been by a factor of ten at least. We are delighted by this but, as with most good things, it has its drawbacks. When research was the spare-time occupation of university teachers or public servants it was unarguable that they should do whatever seemed interesting to them. But now that science is known to be useful, research becomes the duty of university teachers and the full-time occupation of staffs remunerated from public funds; it is a short step from the discovery that science is useful to the belief that science must aim to be useful if it is to enjoy financial support. I hope, indeed I know, that many of us are pleased to smuggle some useless luxury items in amongst our merchandise and, to be fair, the customs officials do not examine our wares too closely, but nonethe-less, we are inflicted with programmes and priorities, and committees which award funds: this is one of the facts of life which it serves no purpose to ignore. It has had a big effect on the development of meteorology.

Meteorology, we should probably all be ready to agree, is the physical science of the atmosphere. I say physical science, for the meteorologist, unlike the oceanographer, does not claim biology: if the oceanographer claims the fish, the meteorologist does not claim the birds and butterflies. But we accept limits neither to the branch of physical science nor to the geographical and vertical extent of the atmosphere which comes within our subject. And yet, having said this, we must admit that most of us who go by the name of meteorologist are really quite restricted in our studies. We have been concerned with weather and climate, we seem to care little for anything but pressure, wind, temperature, and water in its three phases. It is really unfortunate that there is no word in English for weather science and therefore none for the weather scientist. We might have done very well with climatology and climatologist but one doubts if those words will ever completely recover in the minds of physicists from the odium attaching to arid statistics, although there are promising signs. The lack of a specific word has, I say, been really unfortunate in that, for want of another, the general word meteorology has tended to become specialized to weather science with what I believe may be far-reaching and perhaps unwanted results to which I shall refer later. That part of meteorology which is concerned with weather and climate has developed enormously, largely of course because it has been in economic demand and the effort has been made available. Consequently, I, in common with thousands of others who claim the name of meteorologist, am really a weatherman and therefore propose first to say something about the state of weather science.

I came to the subject some thirty years ago and it is natural that I should contrast the present state with that of my early years before World War II. It was, we were fond of saying, an undeveloped subject but it was not at all a new subject. The broad physical background to the phenomena of weather had been filled in before my time and, although there was little research going on, the period was one of consolidation and text-book writing.

If I may sketch the outline very briefly we recall that on the largest atmospheric scale the seasonal world distributions of wind, pressure, temperature and humidity of the troposphere and lower stratosphere were roughly known; the troposphere-stratosphere structure was ancient history; the synoptic climate of the world could be described quite fairly. There was no basic theory of the general circulation but at least the old plausible and facile arguments on steady states had been abandoned and the essential role of variability had been established.

In synoptic meteorology the great advance brought by frontal and air-mass analysis was being consolidated; cyclone and anticyclone structures were well described although basic dynamical theory was missing. Smaller-scale weather phenomena, clouds and fog were understood quite well, vertical instability was well explained with the tephigram and other aerological diagrams in established use; the microphysics of clouds and precipitation which seems such a modern subject was by no means entirely unexplored. But it is true that, compared with the present day, very little research was going on and the large gaps in our knowledge, empirical and theoretical, were being filled in only very slowly.

Since those days and especially since the war with the manifold increase of effort, progress has been very rapid. Taking once more a rapid view, we see the three-dimensional world picture filled in with remarkable and unpredicted detail. The tropopause, earlier thought of as a continuous surface although higher in lower latitudes, is found to break and rise discontinuously, from about 200 mb to 100 mb, in tropical latitudes in association with a sub-stratospheric westerly jet stream. Distinct from this in origin, although sometimes merged with it in the complex play of weather changes, we have the highly fluctuating polar-front systems of middle latitudes marked by further and generally

discontinuous lowerings of the tropopause polewards and associated with further jet streams; at times a separate arctic frontal system will have a similar three-dimensional structure and recent data have revealed a stratospheric westerly jet in the ozone layer on the fringe of the polar darkness. We still await a basic theory of the general circulation but much has been clarified about the mechanisms especially relating to the balancing of momentum, energy and moisture.

In synoptic meteorology there has been a great step forward, perhaps it is excusable if I think it is the most significant step of all. In my early days no one seemed to have anything satisfying to say about the cyclone or anticyclone as a three-dimensional dynamical system; now a fair account can be given in the form of differential equations which can be integrated and even applied to practical weather forecasting. We might also mention the long waves of the atmosphere, the Rossby waves, and the short orographic waves, the Scorer waves, both new discoveries and both well founded on theory. Cloud and precipitation physics has become a highly developed branch of the science with a large volume of empirical data and theoretical studies.

I think we may claim that we have recently lived through a great period in the development of weather science, perhaps this is indeed its golden age; how long, we may ask, will it last ? Perhaps this is a disloyal question, perhaps it is my duty to wave the flag for a science on which the sun will never set, but even so a little sober thinking can do no harm. The development of weather science in this great phase of its history is I think of a fairly obvious kind, the filling in of quantitative data and quantitative theory to what was largely a qualitative science – the transforming of the subject into an exact science. In principle no doubt there is no limit to the degree of exactness which may be aimed at but is it heresy to suggest that beyond a certain stage the old problems will become tedious and the the stage of diminishing returns will be reached ?

Here I would like, in the context of economic usefulness, to draw a distinction between our science and some others. The first and most obvious way in which science and technology are revolutionizing our way of living is by invention, the invention of new materials, new processes, new sources of energy, new tools, new machines. This invention or creation is the distinctive contribution of the physical sciences. The second main influence, on the whole more typical of the biological sciences, is through the control of natural phenomena; we have in mind surgery and medicine, animal husbandry and all forms of scientific agriculture. The thesis would need more careful attention if it were to stand in its own right but I do not think that there can be anything seriously misleading in the view that invention and control are the two tremendous applications of science towards material ends and that both have unlimited potentiality, restricted only by the amount of effort which can be devoted to them. Now I do not see how meteorology can be assigned to either category, now or at any foreseeable time. It may be that weather control will come; if so, the position will be altered as there will be no limit then to what might be thought to be worthwhile. But failing that, the practical economic value of weather science will rest on two other factors, neither invention nor control but what I shall call information and prediction or, as loose meteorological equivalents, climatology and forecasting. It would be a great mistake for the scientist intent upon research to decry or despise the direct value of scientific information to the community. After all, almost the whole of the effort of the International Geophysical Year could be justified as pure empirical geography made attractive to experimental physicists because ingenious apparatus may be employed in collecting the data. I find it difficult to believe that the time will come when the community will not demand weather information for many purposes. Forecasting also has come to stay and I will not here enter into conjecture about its future; I have made my views known elsewhere and in spite of unfailing interest and the weighing of

every piece of evidence that comes my way I cannot yet find positive grounds for expecting a radical advance. Some marginal improvement there must be and perhaps the long-range prediction problem will yet yield to attack, but a really far-reaching change is not at present in view.

It would of course be very easy to speak airily of the possibilities of prediction and express an unbounded confidence in some genius of the future who will solve the problems which today seem so intractable, but such pipe-dreaming is not interesting. As I see it, weather science, through the usefulness of its information and predictions, is an established profession which will demand teaching and scholarship for an unlimited future. Attractive research problems are present now in profusion, this is indeed a time of opportunity which will last perhaps for another generation after which, unless new ways of thinking emerge, further gold may become much harder to win.

But now I wish to return to meteorology in its wider sense, the physical science of the atmosphere, much of which has little or no direct connexion with weather. A great deal of the interest lies in the high atmosphere above the level of 20 km and at these levels the weatherman's problems of wind, temperature, pressure and moisture are not the only ones or even the primary ones. The problems concern the constitution of the atmosphere; absorption, emission and photochemistry; ionization, aurora and airglow; the effects on the atmosphere of solar disturbances and cosmic radiation; the effects of the atmosphere on geomagnetism, on radio telecommunications, on the flight of rockets or the launching and trajectories of artificial satellites. Yet this wider meteorology is not exclusively an extension to the high atmosphere, there are problems of the lower atmosphere which are hardly included in weather science although of course the connexion is here more intimate. A prominent and special problem is that of ozone which spans the lower and higher levels of the atmosphere. Recently the whole question of the constitution of air and rain water, especially in relation to their content of sodium and chlorine, nitrogen and sulphur and other elements and radicals, has begun to receive new attention in many countries; atmospheric radioactivity has become a matter of wide concern, stimulated of course by the effects of man-made radioactivity, especially from weapon trials, but not ending there.

I have not the time, even if I had the knowledge, to survey the progress and importance of these other parts of meteorology; let it suffice to say that they are attracting one way and another a very substantial amount of scientific effort sometimes under the label of meteorology and sometimes not. A very fair question now arises: what do we wish to imply by the word meteorology, shall we allow it to become specialized to the science of weather and climate for which there is no name, or shall we aim to maintain the wide connotation ? First I will list a few indicators of the present tendency towards a dichotomy, for it has gone quite a long way. If we pick up any book with meteorology in the title we shall find that it is in part concerned with weather and climate and usually with little else. We have in this country a *Journal of Atmospheric Physics*: it contains very little weather science. I recently attended a conference on the Arctic atmosphere – it was divided into two parts, one called meteorology, the other the ionosphere. At another conference some workers in atmospheric radioactivity have protested that their studies were not meteorology. Perhaps the most clear pointer of all lies within the International Union of Geodesy and Geophysics which has, amongst others, two Associations, one for Meteorology and the other traditionally for Geomagnetism. Now students of geomagnetism must pay deep attention to the behaviour of the upper atmosphere, especially the ionosphere, and relevant aspects of meteorology have naturally been studied in their Association. But now the proposal is that there should be a new word for those parts of atmospheric physics, they should be called aeronomy, a science distinct from meteorology or at least

assigned to an Association distinct from the Association of Meteorology. Probably only a very few people in this audience are aware that the title 'International Association of Geomagnetism and Aeronomy' has been adopted already: if it is retained the dichotomy in atmospheric science will be crystallized.

But the signs do not all point the one way. The weatherman who is traditionally and intimately associated with aviation is being forced to extend his attention to ever higher levels. The rocket extends the range indefinitely and it is difficult to see how State meteorological services can avoid becoming involved in the high atmosphere. In this country high-atmosphere research by means of rockets is indeed supported by the Air Ministry under the heading of meteorological research. Again, the important and fruitful studies of atmospheric ozone have grown up in relation with meteorology, and atmospheric chemistry and radioactivity are awaking the interest of meteorologists and meteorological services everywhere.

So it seems to me that our science is at an interesting and perhaps critical stage in its evolution with tendencies both to divide into two sciences and to heal the rift which has already occurred. Perhaps you will wonder why I think the question worth attention and this brings me to my last point – the Society.

This is a healthy scientific Society in every way. Our membership is numerically strong and steady; financially we are sound although we share the troubles of inflation and may have to raise our dues; our meetings are well attended and our scientific discussions as lively and stimulating as any in comparable Societies – I think in these respects they are well above the average; our *Quarterly Journal* is respected everywhere and our second periodical, *Weather*, is I think deservedly popular. My own instinct, quite frankly, is to leave well alone but in the case of the Society I do not think we are justified in always letting things take their course merely because they are running smoothly. This Society, as I have insisted elsewhere, is not a club to be run for the pleasure of its members, it is a learned Society with a Royal Charter founded to promote the science of meteorology and we should ask ourselves from time to time whether we are discharging our duty creditably. The answer depends on what we mean by meteorology. If meteorology means weather science we may be properly complacent but if meteorology means the science of the atmosphere we as a Society are giving remarkably little attention to some of its very interesting and important branches. Should we or can we do anything?

We have never positively banned any part of atmospheric physics from the Society but I think we may have slight twinges of conscience for the little effort we have made to hold the subject together. We have called our popular periodical *Weather*; I wonder if it was a good choice. The name does seem to limit the subject rather definitely and I doubt if it is even an attractive name to those broadly interested in science: it is a little commonplace and does not suggest the excitement of modern research and discovery. But that is only a passing thought. The Society has these spacious rooms with excellent facilities for meetings and used, I suppose, for that purpose some 12 days in the 365. At the same time those meteorologists whose main interests are not in the weather seem to have no special Society or meeting place: they are wandering about without a home. It may be that they positively dislike us and do not want to be called meteorologists, but we can at least do more than merely leave the door open, we can invite them in.

Fragmentation in the sciences is not a good thing and the geophysical sciences have much to gain from coming together. The least we can do, if we believe this, is to try to avoid the dichotomy of atmospheric physics into meteorology, equated with weather science, and other studies for which another name will be found. If we can do this I believe our successors will be grateful to us for the enrichment brought to the Society. We have not yet passed the point of no return, the time is ripe for action and the outlook favourable.

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Evidence of a mean meridional circulation in the atmosphere from surface wind observations

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SUMMARY

Surface-wind observations in the Northern Hemisphere are analysed and fields of mean meridional and zonal motion obtained. From these, profiles of mean meridional motion as a function of latitude are deduced. The results are consistent with a direct meridional circulation in low latitudes and an indirect circulation in mid-latitudes, both attaining speeds of 1 to 2 m/sec at the surface. This scheme is found to be consistent with a decrease of upward flux of westerly momentum with height (convergence) in the surface layers in low latitudes and a decrease of downward flux in mid-latitudes.

1. INTRODUCTION

The mean meridional circulation in the atmosphere is defined as the space-time average of meridional velocity. The spatial range is the complete latitude circle and the temporal range must be sufficient to contain a quasi-steady pattern of flow. This circulation is essentially the component of the general circulation in the meridional plane; the general circulation being the basic three-dimensional flow pattern upon which are superimposed the perturbations brought about mainly by various types of instability in the atmosphere.

Until recently, although theory and observation had kept pace in the study of the general circulation, speculation had outstripped both as far as the meridional circulation was concerned. However, during the last decade several workers have been engaged on the problem of assessing the contribution of the meridional circulation to the dynamics of the atmosphere. Information of its existence and function were derived from observational studies of transfer processes and were initiated by Priestley (1949 and later papers) and developed by Starr and White (1951 and later papers) and others. But the results are inconsistent; Priestley (1949) from an angular momentum analysis infers an appreciable direct circulation in 30° to 35°N (0·9 m/sec at 950 mb and 1·8 m/sec at 200 mb), while Starr and White (1951) from a more representative set of observations maintain that any meridional circulation is insignificant.

Riehl and Yeh (1950) have obtained a net surface meridional velocity for latitudes below 40°N, computed from a dense network of ocean surface observations. This together with a similar analysis by Priestley (1950) is the only acceptable observational assessment of the surface component of the circulation. There is an obvious need to extend these analyses by :

(a) incorporating land observations,

(b) extending the field of analysis north of 40°N,

(c) undertaking similar investigations at higher levels.

These extensions are the more important since recent theoretical studies (e.g. Eliassen (1950) and Rogers (1954), have arrived at results supporting the existence of a direct meridional circulation in low latitudes. Even more significant is the result of Phillips (1956) who, by considering a model more close to the actual atmosphere than those used in previous studies, has derived not only a direct circulation in the tropics but an indirect circulation in temperate latitudes.

The first two of the extensions to the work of Riehl and Yeh, and Priestley, mentioned above, have been carried out to derive an accurate estimate of the mean meridional flow at the surface as a function of latitude, and the charts from which this is obtained are presented. In addition some implications of the results as regards stress are considered.

* Now with the Meteorological Office.

2. Data

Velocities are expressed in metres per second with the sign convention: zonal motion, westerlies positive; meridional motion, southerlies positive. The two seasons considered are the Northern-Hemisphere summer (June, July and August) and winter (December, January and February), mean values for the year (12 months) are also derived.

(a) *Surface observations*

All available summaries of surface winds were utilized to derive mean velocity components for five-degree squares over the oceans and a network of points over the land. Data were entered on Northern-Hemisphere charts and isopleths of equal velocity component drawn up. No reference to pressure field was made.

Ocean data were obtained from *Monthly Meteorological Charts of the Oceans* (Meteorological Office, London, 1947-1950) which provide monthly resultant winds based on observations made generally over more than 50 years. The number of observations for each 5-degree (Marsden) square was noted for subsequent reliability assessments.

The problem of obtaining representative means for a network of points over continental areas was aggravated by :

(i) topographic and coastal effects,

(ii) lack of adequate published data over large areas,

(iii) inhomogeneity of periods of observation.

As regards (i), coastal stations and stations situated in rugged country were treated with reserve, and as a result of (ii) and (iii) resort was necessarily made to rather crude data. Interpolation and smoothing were more extensively used in drawing isopleths over land than over the oceans.

The network of land stations used is shown on the Key Chart, the number of years over which the observations extend is entered alongside each station. The majority of the surface winds were worked up from frequency tables. The exception to this was for some stations in the U.S.S.R.; observational material was unavailable and resort was made to climatic charts of mean monthly surface winds given in (i) *Atlas Climatologique de l'Empire de Russie*, St. Petersburg, 1900; (ii) Geophysikalisches Zentral Observatorium. Klima der Union der Sozialistischen Sowjet-Republiken. Teil II, Luftdruck und Wind in der U.S.S.R., Lieferungen 1 and 2. Luftdruck nach Monatsmitteln und Windrichtung in der U.S.S.R., von A. Kaminsky, Atlas, Leningrad, 1932. The period of observation is not given, but winds are represented by barbed arrows for each station. Data from both charts were compared and where the winds for the same station are within 10 degrees of each other the mean direction was taken to be the resultant wind direction and the magnitude taken from the Beaufort force quoted in (i). Values obtained agree well with data for other stations in U.S.S.R. obtained from different sources.

A complete list of sources from which data were extracted is given in the bibliography.

To facilitate subsequent computations the sea-level surface has been regarded as the 1,000 mb level. In many areas the land rises to well over 850 mb (c. 1500 m) and here the contribution to the ' surface ' circulation is zero. Accordingly, as a method of procedure, where 50 per cent or more of the land in a 5-degree square exceeds 1000 m in height, this square was assumed to make zero contribution to the surface air flow.

(b) *Accuracy of observations*

It is necessary to know the reliability of the mean values represented by the isopleths in the charts, and of the space-time means computed for a zonal belt 5 degrees in width.

σ_0 is written for the standard vector deviation of an observed value at a station about the true value, and σ_s for the standard vector deviation about a temporal mean. Assuming that successive wind observations are independent and there is a normal circular distribution of vector deviation about the resultant, then the standard error Σ of a mean velocity component is given by

$$\Sigma = \left(\frac{\sigma_s^2 + \sigma_0^2}{2n}\right)^{\frac{1}{2}}$$

where n is the number of observations.

(i) *Reliability of the charts.* Charts of σ_s above the friction layer over the whole globe were prepared by Jenkinson (1956) who used a statistical relation between σ_s and the mean standard deviation of surface pressure; in addition Mrs. A. H. R. Goldie (Miss N. Carruthers) has compiled unpublished values of σ_s from surface-wind observations over the oceans. Jenkinson's charts were used to extend these ocean values in order to obtain the best spatial distribution of σ_s over land and sea. σ_0 was assumed to be 1 m/sec over land and sea. On this basis and taking a 5-degree square as a working area, values of the standard error are given in Table 1, assuming 10 years' observations are available over the land. However, observations on successive days cannot be considered independent, but if this is true say of every fourth day, Σ is increased by a factor of two. With the exception of land values in winter the charts can therefore be regarded as having a 95 per cent reliability limit of about 0·4 m/sec (about twice the standard error).

TABLE 1. TYPICAL VALUES OF σ_s, σ_0, n and Σ

	Winter		Summer		Year	
	Land	Sea	Land	Sea	Land	Sea
σ_s (m/sec)	8	10	5	6	7	10
σ_0 (m/sec)	1	1	1	1	1	1
n	900	4800	900	4800	3600	19200
Σ (m/sec)	0·19	0·10	0·12	0·06	0·09	0·05

(ii) *Reliability of zonal averages.* Similar considerations to those in section (i) were used to derive the standard error of the mean latitudinal velocity components (Table 2). The variation of the standard vector deviation, σ_s, with latitude is the cause of the variation of standard error with latitude.

TABLE 2. STANDARD ERROR OF MEAN LATITUDINAL VELOCITY COMPONENTS (TO NEAREST 0·05 m/sec)

Latitude (°N)	65	55	45	35	25	15	5
Winter	0·2	0·15	0·15	0·1	0·1	0·05	0·05
Summer	0·15	0·1	0·1	0·1	0·05	0·05	0·05
Year	0·1	0·05	0·05	0·05	0·05	0·05	0·05

3. CIRCULATION NEAR THE SURFACE

(a) *The field of mean velocity component*

Charts of mean velocity component for the Year (Charts 1 and 2), Summer (Charts 3 and 4) and Winter (Charts 5 and 6) are presented without comment. Isopleths are dotted in over high ground for the sake of continuity.

Key Chart. ● Stations, with number of years of observations alongside. ○ Stations, winds taken from climatic charts.

Chart 1. Mean annual zonal velocity component at the surface (m/sec; westerlies positive).

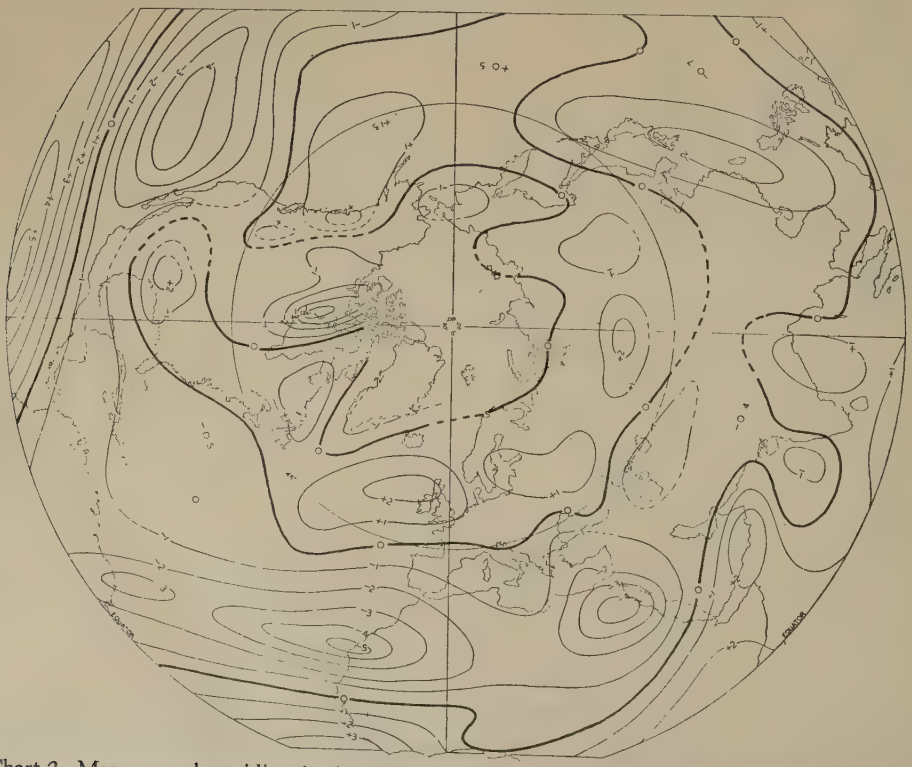

Chart 2. Mean annual meridional velocity component at the surface (m/sec; southerlies positive).

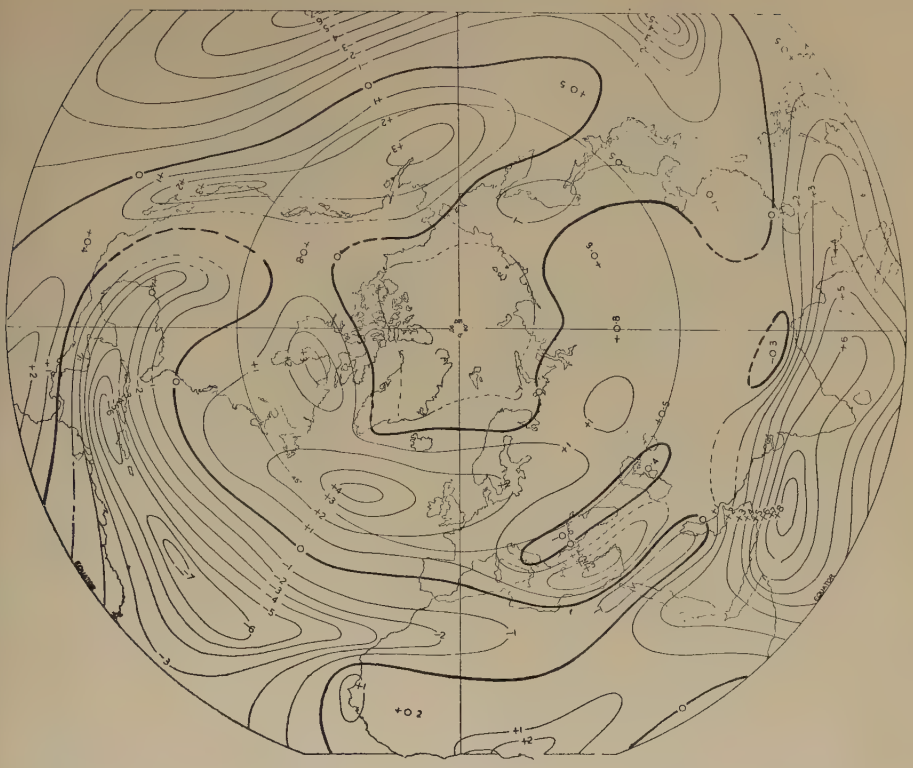

Chart 3. Mean summer zonal velocity component at the surface (m/sec; westerlies positive).

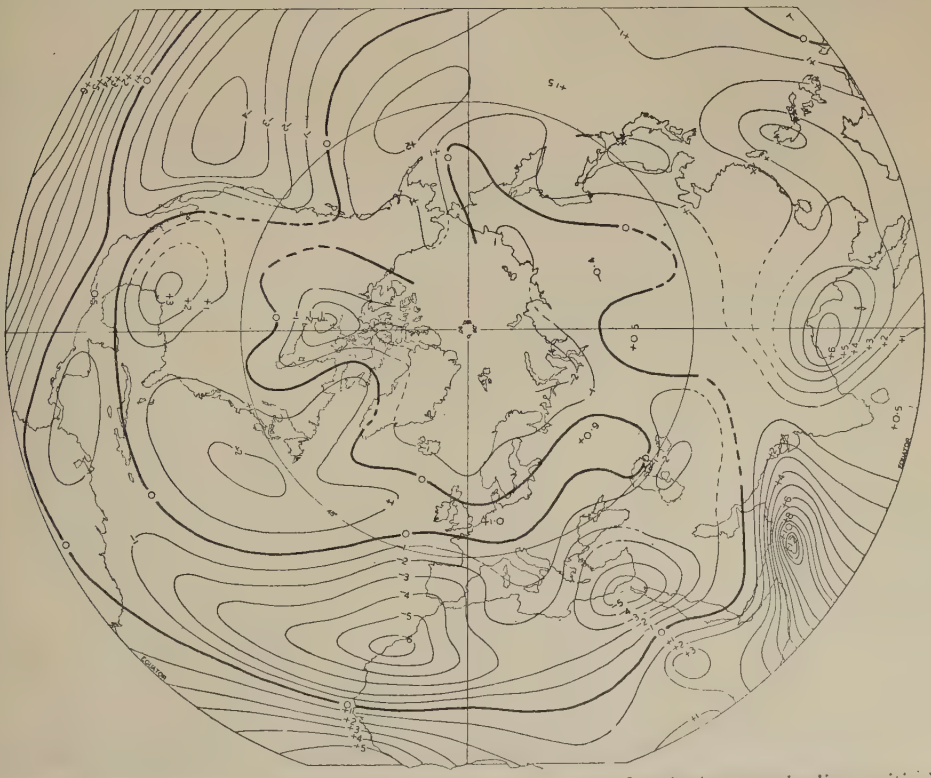

Chart 4. Mean summer meridional velocity component at the surface (m/sec; southerlies positive).

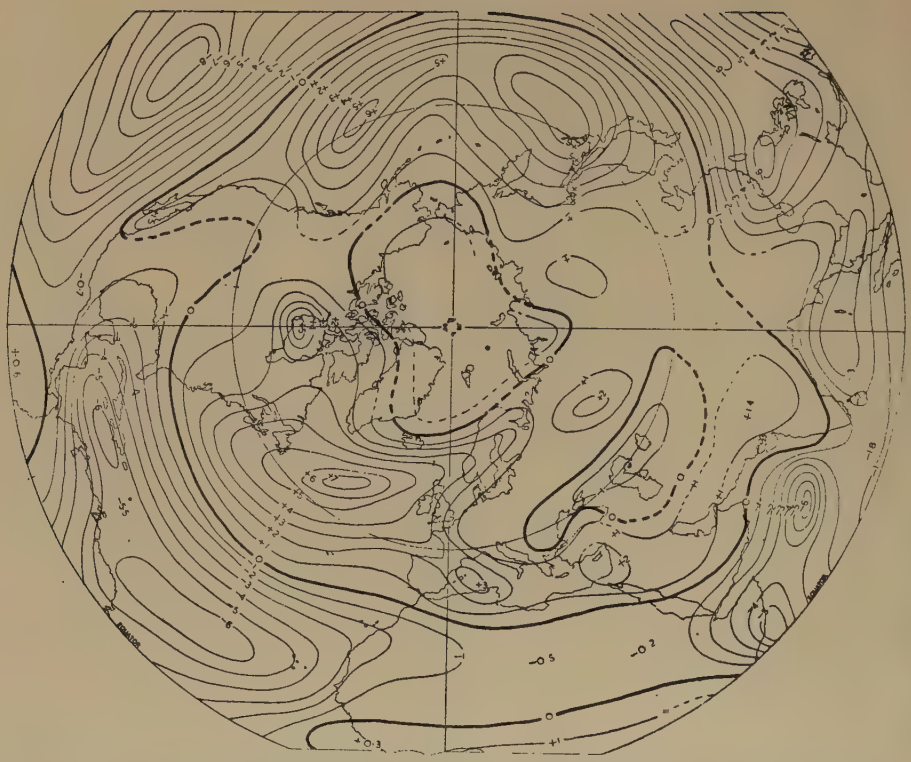

Chart 5. Mean winter zonal velocity component at the surface (m/sec; westerlies positive).

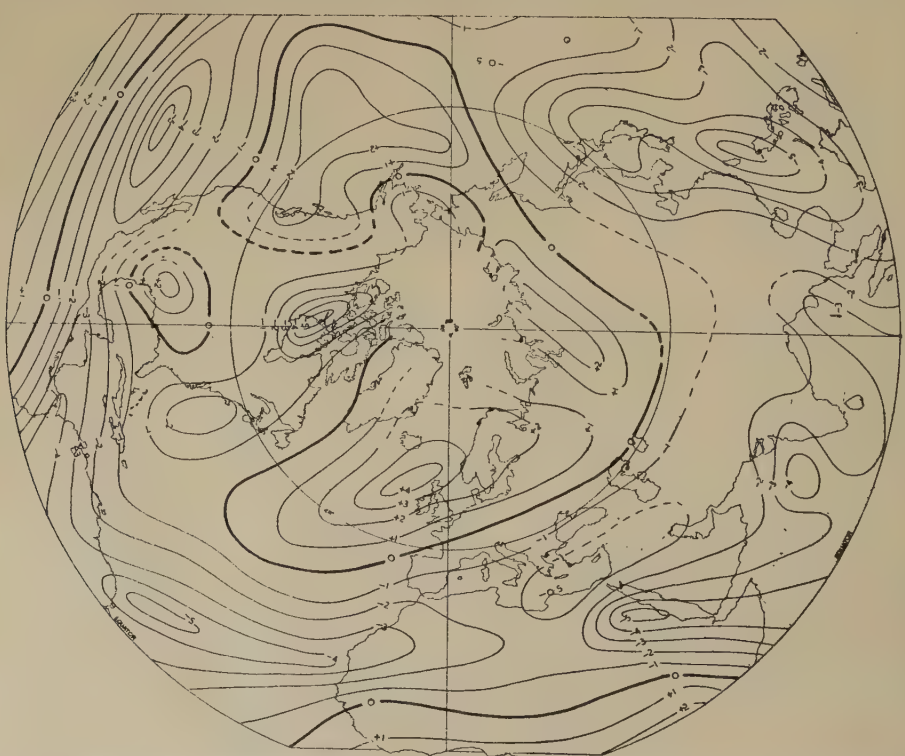

Chart 6. Mean winter meridional velocity component at the surface (m/sec; southerlies positive).

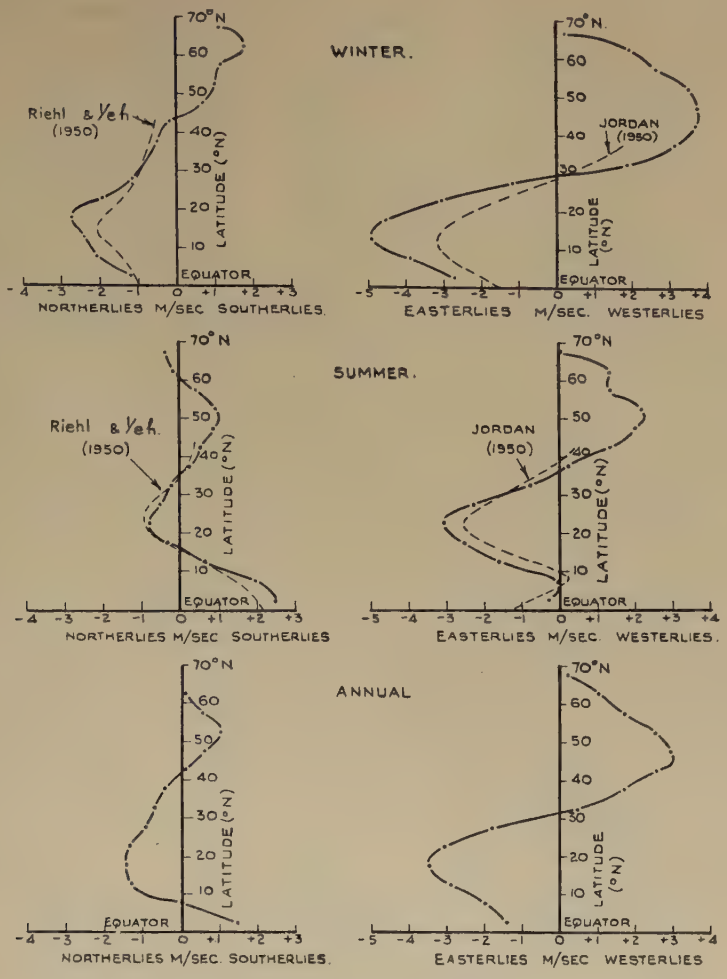

Figure 1. Mean velocity components over the ocean surface.

(b) *Profiles of mean latitudinal velocity component*

(i) *Oceanic values.* Profiles of the mean meridional velocity, $\bar{v}$, and mean zonal velocity, $\bar{u}$, are presented in Fig. 1 and Table 3 and are compared with those of Riehl and Yeh (1950) and Jordan (1950). The sources of data were different, but there is probably considerable overlap; the good agreement is a guide to the reliability. Since neither of the above authors give their spatial distribution of velocity, the significance of the smaller amplitude of their curves cannot be discussed.

(ii) *Land and sea values combined.* When the oceanic curves are compared with the entire circulation over land and sea (Fig. 2 and Table 4), the way in which the former over-emphasize the magnitude of the components is revealed. Comparison of the $\bar{u}$ profile (Fig. 1) with the profile of mean pressure $\bar{p}$ along a latitude circle (Fig. 3, U.S.W.B., 1952) shows that the changeover from tropical easterlies to mid-latitude westerlies approximately coincides with the high pressure ridge in summer and winter, but in winter the corresponding value of $\bar{v}$ shows a cross-ridge northerly flow of about 1 m/sec. Riehl and Yeh also noted this and suggested it was the result of equatorward mass transport in the lower levels at the eastern ends of the ' subtropical cells ' exceeding the return flow at their western margins. The encroachment of the Southern-Hemisphere circulation system north of the equator is a feature of both the summer and annual profiles.

TABLE 3. MEAN VELOCITY COMPONENTS OVER THE OCEAN SURFACE (metres/sec)

Latitude (°N)	Winter		Summer		Annual	
	Meridional	Zonal	Meridional	Zonal	Meridional	Zonal
65-70	+ 1·1	+ 0·2	− 0·3	+ 0·1		+ 0·2
60-65	+ 1·8	+ 1·9	− 0·2	+ 1·3	+ 0·1	+ 1·1
55-60	+ 1·1	+ 2·6	+ 0·4	+ 1·4	+ 0·6	+ 1·7
50-55	+ 1·0	+ 3·5	+ 0·9	+ 2·2	+ 1·0	+ 2·6
45-50	+ 0·7	+ 3·5	+ 0·9	+ 2·0	+ 0·6	+ 2·9
40-45	− 0·2	+ 3·7	+ 0·9	+ 1·4	+ 0·1	+ 2·5
35-40	− 0·5	+ 3·1	+ 0·5	+ 0·2	− 0·4	+ 1·7
30-35	− 0·8	+ 1·9	+ 0·3	− 0·8	− 0·7	+ 0·5
25-30	− 1·1	− 1·1	− 0·5	− 2·3	− 0·9	− 1·8
20-25	− 1·9	− 3·0	− 0·8	− 3·0	− 1·3	− 3·1
15-20	− 2·7	− 4·5	− 0·3	− 2·5	− 1·4	− 3·5
10-15	− 3·0	− 4·9	+ 0·6	− 1·6	− 1·3	− 2·9
5-10	− 2·1	− 3·9	+ 2·0	0	0	− 2·0
0- 5	− 1·2	− 2·7	+ 2·5	− 0·3	+ 1·5	− 1·4

TABLE 4. MEAN VELOCITY COMPONENTS OVER THE SURFACE, LAND AND SEA COMBINED (metres/sec)

Latitude (°N)	Winter		Summer		Annual	
	Meridional	Zonal	Meridional	Zonal	Meridional	Zonal
65-70	+ 0·4	+ 0·4	− 0·7	0	− 0·2	+ 0·1
60-65	+ 0·5	+ 1·1	+ 0·4	+ 0·6	0	+ 0·8
55-60	+ 0·6	+ 1·8	+ 0·1	+ 1·2	+ 0·4	+ 1·4
50-55	+ 0·7	+ 2·0	+ 0·5	+ 1·4	+ 0·7	+ 1·6
45-50	+ 0·2	+ 1·9	+ 0·5	+ 1·2	+ 0·3	+ 1·5
40-45	− 0·4	+ 2·0	+ 0·1	+ 0·8	− 0·1	+ 1·3
35-40	− 0·5	+ 1·9	− 0·1	+ 0·3	− 0·4	+ 0·9
30-35	− 0·7	+ 1·2	− 0·4	− 0·3	− 0·6	+ 0·3
25-30	− 1·3	− 0·6	− 0·7	− 1·4	− 1·0	− 1·2
20-25	− 2·1	− 2·1	− 0·4	− 1·6	− 1·4	− 2·0
15-20	− 2·6	− 3·5	0	− 1·5	− 1·3	− 2·5
10-15	− 2·7	− 4·0	+ 0·6	− 1·0	− 1·0	− 2·3
5-10	− 1·6	− 3·0	+ 1·9	− 0·3	+ 0·1	− 1·6
0- 5	− 0·7	− 2·0	+ 2·7	0	+ 1·3	− 1·2

(c) *Divergence near the surface*

Profiles of mean meridional velocity divergence given in Fig. 3 were computed from the formula

$$\mathrm{div}_2 V \equiv \mathrm{div}\, \bar{v} = \frac{\partial \bar{v}}{r\partial \phi} - \frac{\bar{v}}{r}\tan\phi$$

where $\mathrm{div}_2 V$ is the horizontal velocity divergence, ϕ the latitude and r the radius of the earth (latitude steps of 5 degrees were used in the computations). Compared with similar profiles of Riehl and Yeh (1950) the magnitudes are similar, but the importance of using land data is shown clearly. A zone of maximum divergence lies well on the equatorward side of the subtropical high-pressure belt, but another equally strong divergence belt is located further north. A zone of maximum convergence occurs at about 60°N and is most noticeable in summer where it coincides with the low-pressure belt. The northern belt of divergence varies little in magnitude (about 1×10^{-6} sec^{-1}) or position (40°N-50°N) but the belt further south decreases in magnitude and moves northwards from winter to summer. This may be a reflection of the different mechanisms driving the temperate and tropical circulation systems. The tropical meridional motion is probably the result of a direct solenoid drive, the field of motion moving north in summer but the thermal gradients becoming weaker. The low-latitude convergence zone maintans its strength throughout the year.

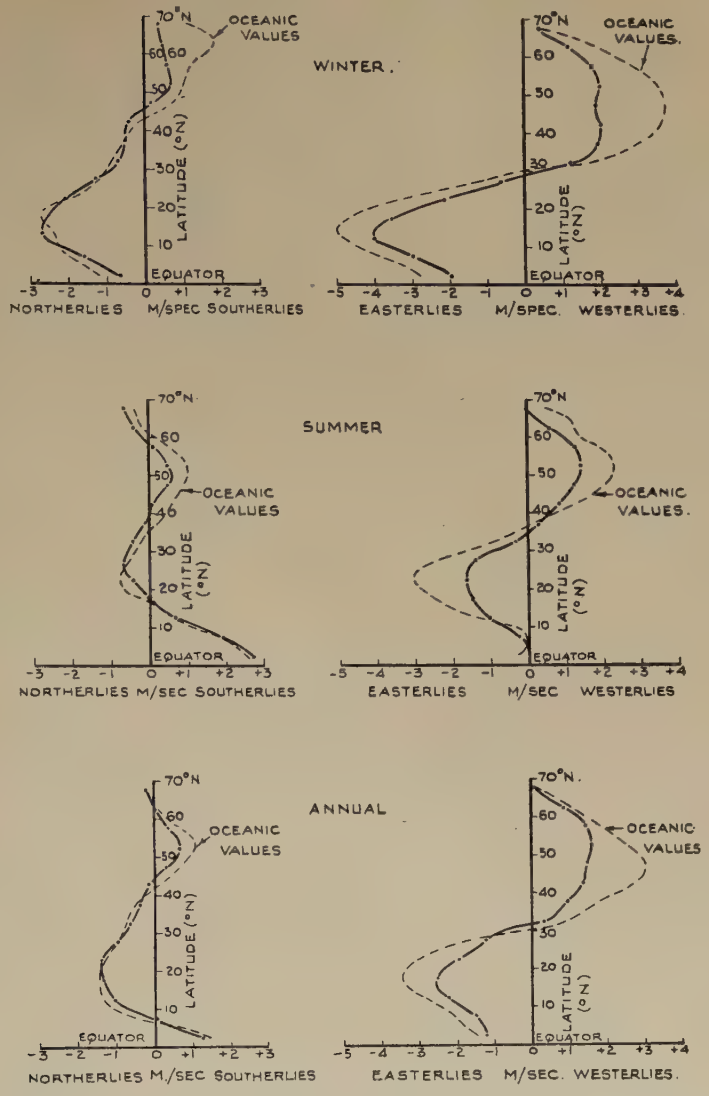

Figure 2. Mean velocity components over the surface, land and sea combined.

4. Some stress considerations

Recently Sheppard (1954) has outlined a theory of the pattern of vertical flux of momentum in the atmosphere, and has used climatological values of wind and pressure gradient at the surface over the oceans to obtain the order of magnitude of the convergence of vertical flux (i.e., the rate of change of shearing stress with height) in the surface layers of the atmosphere in order to confirm his scheme. A similar analysis of derived $\bar{v}$ and $\bar{u}$ profiles will be undertaken below and the results compared with those of Sheppard.

The equation of meridional motion for horizontal rectangular axes x (eastward) and y (northward) is

$$\frac{dv}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + \frac{1}{\rho}\frac{\partial \tau_{xz}}{\partial z} + \frac{1}{\rho}\frac{\partial \tau_{xy}}{\partial y} \quad . \quad . \quad . \quad (1)$$

where z is the vertical axis, ρ the density, l the Coriolis parameter and τ_{xy} and τ_{xz} are the components of the zonal shear stress in the y and z directions. If we consider

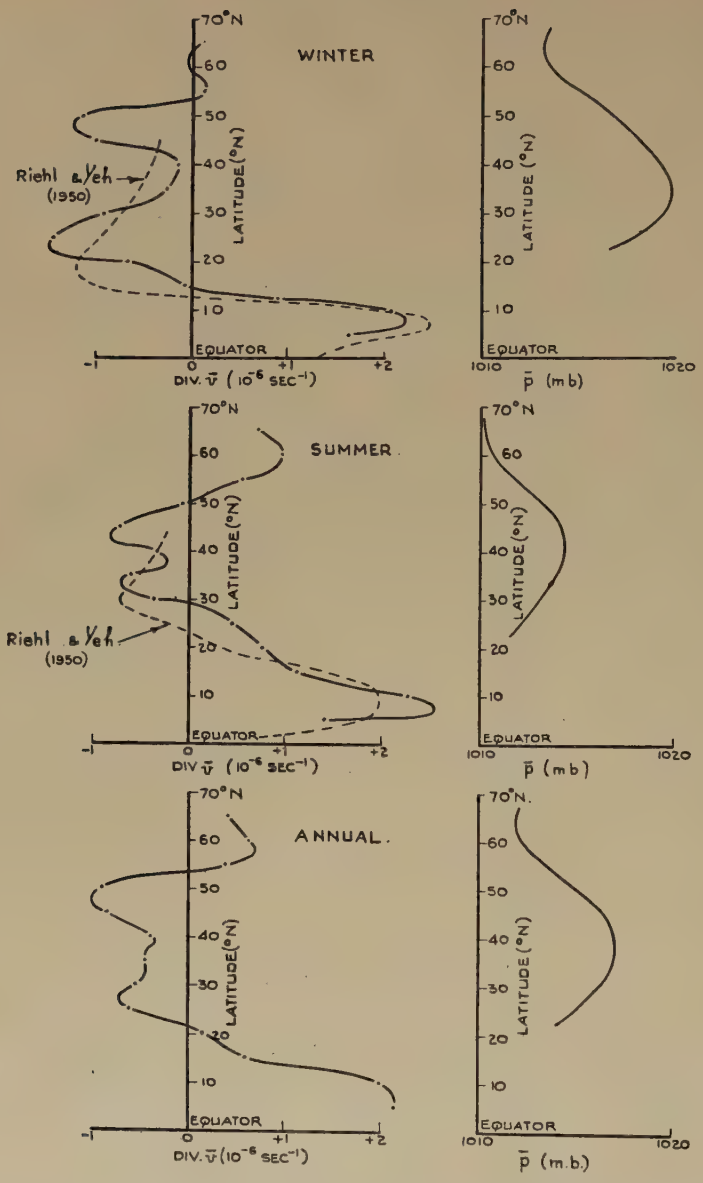

Figure 3. Mean divergence near the surface, and mean sea-level pressure profiles.

a period of time over which the accelerations of the air may be neglected, $\frac{\partial v}{\partial t} = 0$, and assume the other terms in $\frac{dv}{dt}$ to be negligible when averaging along a circle of latitude, then

$$-\rho l\bar{v} = \frac{\partial \bar{\tau}_{xz}}{\partial z} + \frac{\partial \bar{\tau}_{xy}}{\partial y} \quad . \quad . \quad . \quad . \quad (2)$$

But $\frac{\partial \bar{\tau}_{xy}}{\partial y} = -\rho \frac{\partial}{\partial y} \overline{u'v'}$ where u' and v' are the deviations from the local mean values u and v. The maximum value of $\frac{\partial}{\partial y} \overline{u'v'}$ is about 5 m^2 sec^{-2} (10° lat)$^{-1}$ between 35°N and 45°N (Starr 1951), this gives a maximum $\frac{\partial \bar{\tau}_{xy}}{\partial y}$ at the surface of 0·005 dynes cm^{-2} (100 m)$^{-1}$.

The order of magnitude of $\frac{\partial \bar{\tau}_{xz}}{\partial z}$ is 0·1 dynes cm^{-2} (100 m)$^{-1}$ and thus the lateral stress may be neglected *near the surface*. Therefore

$$\frac{\partial \bar{\tau}_{xz}}{\partial z} = -\rho l \bar{v} \qquad (3)$$

Values of $\frac{\partial \bar{\tau}_{xz}}{\partial z}$ are given in Table 5 together with $\bar{\tau}_{xz}$ the shearing stress in the westerly direction (Priestley 1950).

TABLE 5. FLUX OF WESTERLY MOMENTUM AT THE SURFACE, $\bar{\tau}_{xz}$ (AFTER PRIESTLEY), AND CONVERGENCE OF FLUX WITH HEIGHT, $\frac{\partial \bar{\tau}_{xz}}{\partial z}$

Latitude (°N)		5	15	25	35	45	55
Winter	$\frac{\partial \bar{\tau}_{xz}}{\partial z}$ (dyne/cm^2 / 100 m)	− 0·02	− 0·13	− 0·14	− 0·07	− 0·03	+ 0·09
	$\bar{\tau}_{xz}$ (dyne/cm^2)	+ 0·5	+ 0·7	+ 0·2	− 0·45	− 0·85	− 0·65
Summer	$\frac{\partial \bar{\tau}_{xz}}{\partial z}$ (dyne/cm^2 / 100 m)	+ 0·04	+ 0·01	− 0·05	− 0·03	+ 0·04	+ 0·06
	$\bar{\tau}_{xz}$ (dyne/cm^2)	− 0·25	+ 0·5	+ 0·3	− 0·05	− 0·15	− 0·3
Year	$\frac{\partial \bar{\tau}_{xz}}{\partial z}$ (dyne/cm^2 / 100 m)	+ 0·01	− 0·06	− 0·10	− 0·06	0	+ 0·09
	$\bar{\tau}_{xz}$ (dyne/cm^2)	+ 0·1	+ 0·2	+ 0·4	− 0·35	− 0·45	− 0·5

Three conclusions may be drawn from Table 5 :

(i) Between about 15°N and 35°N the values are consistent with a decrease of upward flux of westerly momentum with height in the surface layers.

(ii) North of about 35°N the change in scheme is well marked with a strong suggestion of *increase* with height of *upward* flux of momentum, i.e. in these latitudes a decrease of downward flux of momentum with height.

(iii) The increasing depth of easterlies southwards and their relative constancy with height in the lower layers tends to diminish $-\frac{\partial \bar{\tau}_{zx}}{\partial z}$, but the change of sign in low latitudes in summer reflects again the encroachment of the Southern-Hemisphere regime. Increasing convective activity in very low latitudes causes the concept of the friction layer to be less valid.

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A synoptic study of West African Disturbance Lines

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Summary

The paper gives an account of Disturbance Lines (D/Ls) observed on the charts of the forecasting office of the Ghana Meteorological Department at Accra in 1955. The progress of one of the larger D/Ls is shown to indicate the extensive area affected in some cases. The origins of D/Ls, their movement, distribution, and other characteristics, are described. The importance of the phenomenon in the local weather is shown by the proportion of the monthly rainfall due to D/Ls at four Ghana stations. Emphasis is placed on the nature of the D/L as a disturbance which moves through the atmosphere, transferring convective energy to other forms, notably horizontal kinetic energy, in an ordered manner.

1. Introduction

Characteristics of West African Disturbance Lines (D/Ls) have been clearly described by Hamilton and Archbold (1945). The aim in this paper has been to show how D/Ls are distributed, both geographically and in time, and to give some indication of the varieties of manifestations associated with them. Any phenomenon of the thunderstorm or squall type which appears to move across the area with a degree of regularity has been included. Lines varying in length from 10 to 500 mi, and with lives from 6 to 60 hr, are all considered together. (This includes, therefore, those referred to by Hamilton and Archbold as ' intermediate,' p. 254). In some cases there is no direct evidence on the chart that a *line* of disturbed weather exists, but where thunderstorms with squalls from the same direction are seen to move consistently over a succession of stations which are not necessarily in a straight downwind line, synoptic experience and visual observations suggest that the phenomenon is similar to the larger disturbance. In the following sections those manifestations of a D/L which give rise to the worst weather (thunderstorm, moderate or heavy rain, squall, poor visibility, low cloud base) are referred to as main effects. The less vigorous manifestations which generally occur to the rear of the main effects (slight rain, cool dry surface air, small amounts of low cloud, much thick medium cloud, relatively high surface pressure) are referred to as post-D/L conditions or effects.

An attempt has been made to record each D/L which was observed within the area from the equator to 20°N, and 15°E to the west coast, during 1955. The analysis has been made from the working synoptic charts of the forecasting office of the Ghana Meteorological Department at Accra. These comprise eight charts daily at the main and subsidiary synoptic hours. The charts cover the area from 20°E to 20°W, 7°S to 22°N, scale 1 : 7,500,000. Charts covering a somewhat larger area, extending east to the Nile valley, and further north and south, are used for two of the eight synoptic hours (06 and 12). Fig. 1 shows the main geographical features of the area and the positions of observing stations. The distribution of observations is, however, very uneven, both in space and time. Many station are not operated for all 24 hr, and wireless reception is frequently poor during the early hours of the morning so that the 00, 03 and 21 GMT charts have considerably less than half the numbers of plotted stations compared with the others. Few weather reports from ships are obtained and little precise information is available about D/Ls over the sea. There is a regular series of pilot-balloon ascents with the usual limitation of visual observations, but upper-air charts are necessarily rudimentary. Radio soundings are made at the places shown in Fig. 1 but only those at Dakar and Lagos are made daily (once each).

Figure 1. West Africa, showing synoptic stations. Stations making radio-soundings of temperature, pressure and humidity and radar wind determinations are underlined.

2. A TYPICAL DISTURBANCE LINE

The large area which may be affected by an individual D/L is illustrated in Fig. 2, which shows the positions, at 3-hr intervals, of the D/L of 26-28 May 1955. It is typical of the larger lines. It can be placed to within about 30 mi for most of the period between 06 GMT, 26 May and 03 GMT, 28 May. Over Ghana the position can be even more accurately specified.

Fig. 3 shows the rainfall which resulted over Ghana. North of about 8°N the distribution appears random, amounts fluctuating between nil and $1\frac{3}{4}$ in over as little as 20 mi. South of this latitude amounts were fairly large over the Togoland hills, with a marked decrease immediately to the west and another minimum in the valley of the River Pra. Maxima near Obuasi and Mampong may have been related to the higher ground in these regions.

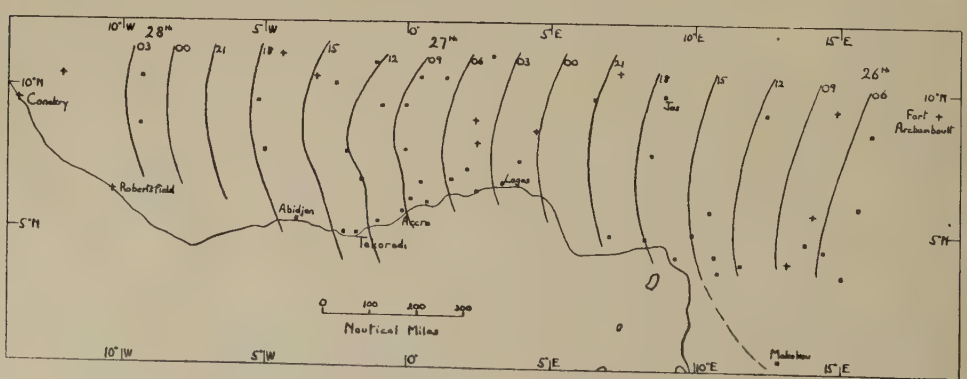


Figure 2. Position of D/L 26-28 May at 3-hr intervals. Main effects indicated ■. Report of post-D/L conditions indicated +.

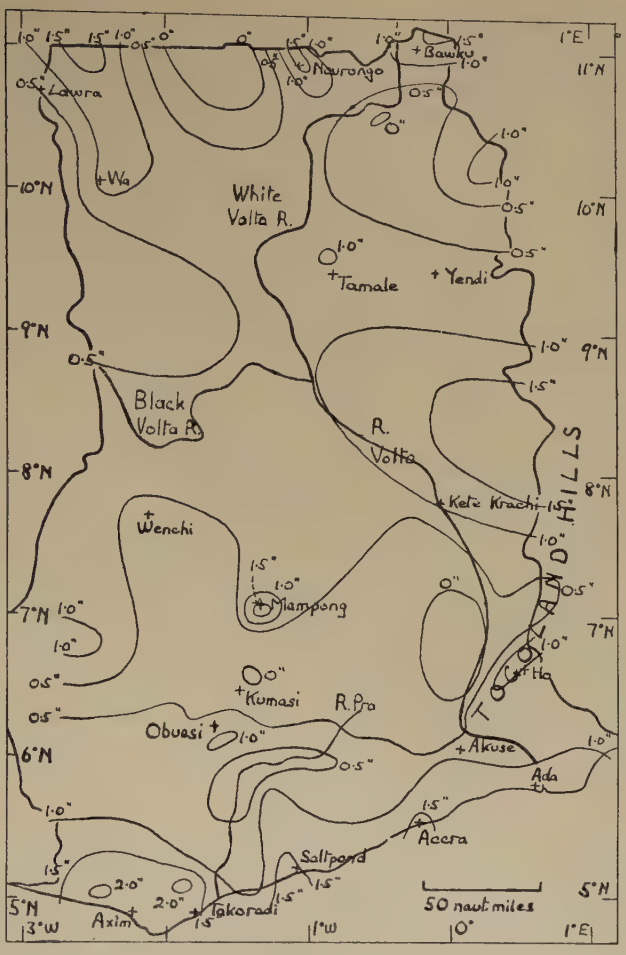


Figure 3. Rainfall in inches for D/L over Ghana 27 May 1955.

3. Distribution

The number of D/Ls within certain ranges of length, and the numbers traversing certain distances, are shown, by months, in Table 1. In this table figures in brackets under ' Length ' give the numbers of occasions where the length was probably greater than was positively indicated. The uncertainty was due either to one end being over the sea, or to lack of observations inland. Those in brackets, under ' Distance traversed,' give the occasions where, if indications of post-D/L effects had been included, the entries would have been in one of the columns to the right (over 2000 mi in the cases of the last column).

Table 2 relates the data given in the two sections of Table 1, in terms of yearly totals. The figures in both these tables must be regarded as giving very general indications only. The analysis is inevitably strongly biased towards areas near Ghana. The larger number of shorter D/Ls shown for months around April and October is largely owing to the greater density of observing stations near the coast.

Fig. 4 shows the tracks of the centre of each D/L during June 1955. Other months would be generally similar but could be indicated less completely as the D/L belt would, in most cases, be partly over the sea.

TABLE 1. SUMMARY OF LENGTHS OF, AND DISTANCES TRAVERSED BY, DISTURBANCE LINES IN 1955

	Length (naut. mi)					Distance traversed (naut. mi)				
Month	< 50	50-150	150-250	250-350	> 350	< 500	500-1000	1000-1500	1500-2000	Total
Jan.	2 (2)	3 (2)	1			6 (1)				6
Feb.	1 (1)	3	2	1		7				7
March	1 (1)	14 (8)	4 (1)	6 (3)	1	20 (4)	5	1 (1)		26
April	2 (1)	17 (6)	6	7 (1)	1	27 (7)	5 (1)	1		33
May	1 (1)	2 (1)	8 (1)	5 (1)	3	11 (8)	5 (1)	2 (2)	1	19
June		4 (2)	1	12 (1)	2	14 (6)	11 (6)	3 (2)	1 (1)	29
July		5	9	12		14 (8)	6 (3)	3 (2)	3 (1)	26
Aug.		3	15	10	3	16 (10)	13 (5)	2 (2)		31
Sept.		4	12	18	4	19 (6)	9 (4)	7 (2)	3 (1)	38
Oct.		17 (2)	13	9	5	38 (13)	5 (1)		1	44
Nov.	1 (1)	10 (1)	2	3	1	17 (3)				17
Dec.	1 (1)	2	2	2		6	1			7
Total	9	84	85	85	20	195	60	19	9	283

TABLE 2. TOTALS OF OCCASIONS WHERE D/Ls OF LENGTHS SHOWN TRAVERSED THE DISTANCES SHOWN

Length (naut. mi)	Distance traversed (naut. mi)			
	< 500	500-1000	1000-1500	1500-2000
< 50	9			
50-150	75	9		
150-250	59	25	1	
250-350	41	24	14	6
> 350	11	2	4	3

4. ORIGINS

(a) *Diurnal development*

There are many occasions when the D/L is a development from afternoon or evening convection. Frequently these D/Ls appear to move away from areas of high ground. Several arise from either the northern or southern parts of the Togoland and Dahomey hills (see Fig. 1) and the high ground 200 mi from the coast of Liberia, Sierra Leone and French Guinea. Experience at Lagos shows they develop from the Oshogbo hills.

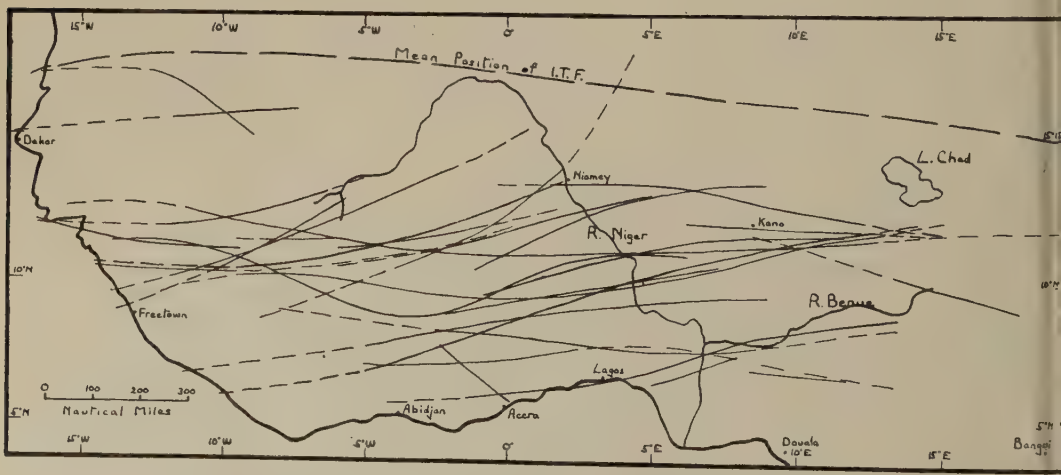


Figure 4. Paths of centre points of D/Ls in June 1955. A dashed line signifies post-D/L reports only.

amilton and Archbold give evidence for those arising from the Jos plateau and the ameroons. Some almost certainly start in the Marra mountains (13°N 24°E). Disturances originating in this way may occur less than 100 mi from the Inter-tropical Front .T.F.), i.e., within Hamilton and Archbold's zone B. When the I.T.F. lies north of bout 18-20°N, in the middle months of the year, several D/Ls appear to develop from nvection over, or to the north of, Northern Nigeria or north French Equatorial Africa nd also over the Upper Niger area.

In one or two cases it was possible to obtain a maximum figure for the rate at which D/L lengthens by observing the development of a line in its very early stages. Also n some occasions D/Ls 300 mi long were evident on the 06 GMT chart when none was pparent 12 hr previously. Both approaches indicate a rate of growth of the order of 5 kt indicating that each end may extend at about 10-15 kt.

(*b*) *Nocturnal developments*

Hamilton and Archbold state (p. 255) that D/Ls which are apparent in the morning an usually be traced back to the charts of the previous evening. While this is frequently o, there was a number of occasions when the D/L was not apparent during the evening p to midnight, but was evident on charts in the latter half of the night or during the norning. Observations are generally less numerous during the night and these D/Ls nay have developed in the 'normal' way. It nevertheless appears that D/Ls may not lways develop from normal diurnal convection (e.g., 29-30 Jan. over Ivory Coast).

(*c*) *Other origins*

On a relatively small number of occasions thunderstorms in the dry air to the north f the I.T.F. take place a few hours before the start of a D/L in the area to the south f the I.T.F. The difference in time and the relative positions of these occurrences are, n some cases, consistent with the speed and direction of the wind at 10,000-20,000 ft. This suggests that the establishment of conditions favourable for the start of a D/L may e affected by advection of air from outside the D/L belt. It is, however, emphasized hat this is a comparatively infrequent occurrence and Hamilton and Archbold's statement p. 253) that the air in the D/L belt is not greatly modified by air from other sources is till regarded as generally true.

5. Movement

(*a*) *Speed*

A wide range of speeds appears, from 17 to 45 kt. Simple averages of speeds of all D/Ls in January and June 1955, weighted for their duration, give 23 and 30 kt respectively. As all D/Ls recorded in January were relatively near to the I.T.F. the lower speed might be expected in view of the relative shallowness and weakness of the easterly winds in this region. Those D/Ls which traversed paths more than 1,000 mi long varied in average speed from 22 to 36 kt with an average between 30 and 31 kt (average taken over the whole year). There is no obvious diurnal variation in speed.

There are many indications that D/Ls move faster over the region between about 9°N and 17°N, and 10°E to 7°W, than over most other areas. Those D, Ls with a marked southerly component to their movement move more slowly than those with the more normal westward movement.

(b) *Speed compared with upper winds*

For the first four months a record was made of the mean 10,000 ft wind in the area up to 300-600 mi ahead of each D/L and for a similar distance in the rear, when D/Ls were near Ghana. At 10,000 ft the component from the east is generally little less than that at any level below 30,000 ft as may partly be seen by reference to Fig. 9. In only 2 cases out of 35 studied was the wind ahead greater than the speed of movement of the D/L and one of these rested upon a single observation. Generally the wind speed at 10,000 ft ahead of the D/L was 10-20 kt less than the speed of the D/L itself. There was generally a gradual increase from the west side to the east and the 10,000 ft wind some distance in the rear (order 300 mi or more) often approximated to the speed of the D/L.

Six cases were studied where the Lagos radar wind ascent was made within 100 mi of the rear of a D/L. With the exception of a 5,000 ft layer in one case, the westerly component of all the winds to 500 mb was not greater than that of the D/L and in many cases it was distinctly less. Thus there is considerable evidence to show that many D/Ls move at a greater velocity than that of the upper winds, at least in the lower part of the atmosphere.

(c) *Direction*

Disturbance Lines were seen to move in any direction between south and north-west. The normal direction of movement of the larger ones is west but many of intermediate size move in a west-south-westerly direction. Movements differing much from these are almost entirely confined to the smaller phenomena, or to the trailing south end of the larger ones. The trailing end may lie almost east-west and it may move south over a considerable distance. On occasions this part appears to have a longer life than that of the normal westward-moving part. Since the trailing end tends to die out at its south-east or east end, the effect is of a line orientated north-west to south-east, yet moving from east to west.

Developments on 9-10 September were unusual, but not unique. At 18 GMT on 9th a D/L was about 70 mi east of Yendi, as shown in Fig. 5. At Yendi thunder and lightning were reported at 19 GMT but no rain or squall occurred during the evening. Normally the main effects would have been expected there between 20 and 21 GMT. It appears that the D/L moved on in latitudes further north, and the south end trailed, and moved south, lying in an east-west direction. It gave thunderstorms at Yendi and Tamale, 50 mi apart, at exactly the same time, 0445 GMT the next morning, Wa having been affected one hour earlier.

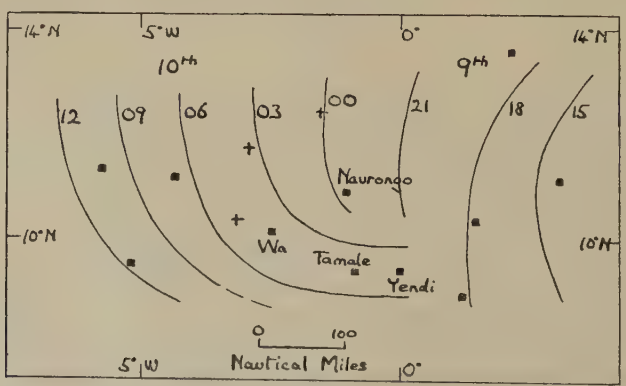


Figure 5. Position of D/L 9-10 September. Notation as in Fig. 2.

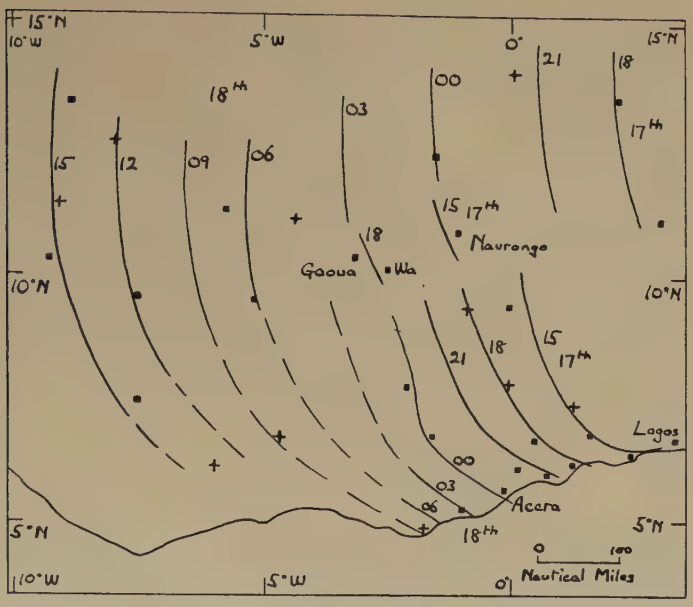


Figure 6. Positions of D/L 17-18 September. Notation as in Figure 2.

(*d*) *Movement in relation to other Disturbance Lines*

The two following cases show that two D/Ls in different latitudinal belts appear to be little affected by each other and that the passage of a D/L may inhibit a line following closely behind it.

During the afternoon of 17 Sept. a small D/L moved across the Ghana Northern Territories affecting Navrongo, Wa and Gaoua as shown in Fig. 6. A second, which either developed over the Dahomey – Togoland high ground, or may have been a continuation of an earlier one, moved slowly to affect most of that part of Ghana not affected by the first. A third, over North Nigeria at 06 GMT on 17th, moved rapidly, without affecting Navrongo and Wa, and appeared to link up with the rather degenerate one in the south by about 06-12 GMT on 18th. The combined disturbance moved over the remaining parts of West Africa to the west coast between 6°N and 12°N.

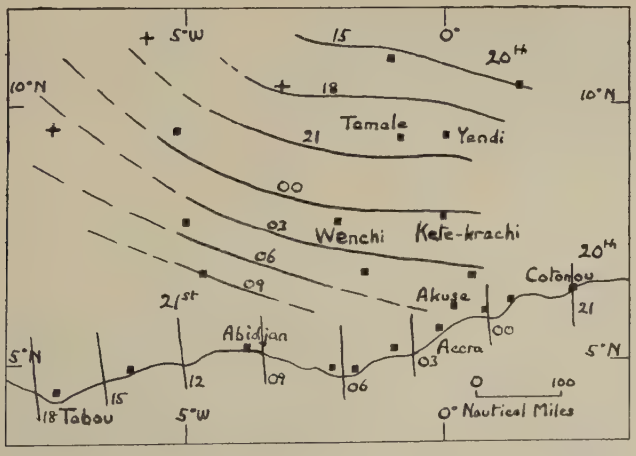


Figure 7. Positions of D/Ls 20-21 March. Notation as in Figure 2.

At 21 GMT on 20 March a D/L was affecting Cotonou as shown in Fig. 7. Thi may have been the northern end of a line which was moving west in a normal manne over the sea, though earlier evidence is inconclusive. It moved at a steady speed of 30 k reaching Tabou at 1719 GMT the following day. No reports were received from station inland of effects from this D/L, except for Akuse (6°N 0°E) about 30 mi from the coast During the afternoon of 20 March an east-west D/L moved south, affecting Tamal and Yendi at about the same time, and Kete-Krachi and Wenchi within 35 min of each other. A thunderstorm and squall at Akuse at 0455 GMT may have been due to this second D/L although the first had affected it about three hours previously.

Major D/Ls do not generally follow rapidly after each other. The following figures showing times of passage of lines at Accra in March, indicate the most rapid succession observed there in 1955 : 15th at 1300 GMT, 17th at 0030, 17th at 1930 (no effects beyond Accra), 18th at 2320, 21st at 0145.

(e) *Movement over the sea*

Hamilton and Archbold consider the movement of existing D/Ls over the sea, (pp. 253 260). Over certain areas it is possible to obtain positive reports of the passage of a D/L after it has moved over the sea. Thus vigorous D/Ls can often be traced from the coast near the equator through the islands of Sao Thome and Principe (a sea distance of 150- 200 mi) with their effects undiminished. In some cases there is very good evidence for continuity between a D/L over equatorial Africa and an equally vigorous D/L over Ghana (a sea distance of about 700 naut. mi).

6. Other characteristics

The south end of a D/L on occasions merges with an area of rain, which may be slight to heavy, but with no squalls or thunderstorms. This extends up to 200 mi south from the latitude of the end of the D/L and several hundred miles east from the longitude of the D/L. Not all D/Ls, however, are associated with this type of rain area, often simply referred to as 'monsoon rain,' and not all the rain areas south of the D/L belt are actually associated with D/Ls. A similar rain area, but with sporadic outbreaks of thunderstorms, also occurs rather less frequently, and over a less extensive area.

Disturbance Lines frequently die out leaving a belt of medium cloud from which slight rain falls for a time. Three other types of ending may be significant :

(a) The D/L moves into an area where widespread convection is about to take place, or is already doing so. This occurs frequently over parts of equatorial Africa near the coast, i.e., 5°S to 5°N, approximately 15°E to the coast.

(b) The D/L moves into an area through which another D/L has passed within a period of about 24 hr or less.

(c) The D/L moves into an area where there is already extensive layer-type cloud, usually with rain.

At times, secondary effects, of the thunderstorm or squall type, appear within 6-12 hr of the passage of the main effects of a D/L.

7. Some geographical considerations

The Disturbance Line is clearly a major factor in the weather of West Africa but D/Ls do not occur as frequently in some areas as in others. Fig. 8 shows the rainfall for each month of 1955 taken from the autographic rainfall records for Accra, Axim, Kumasi and Navrongo, and the proportion of each due to D/Ls.

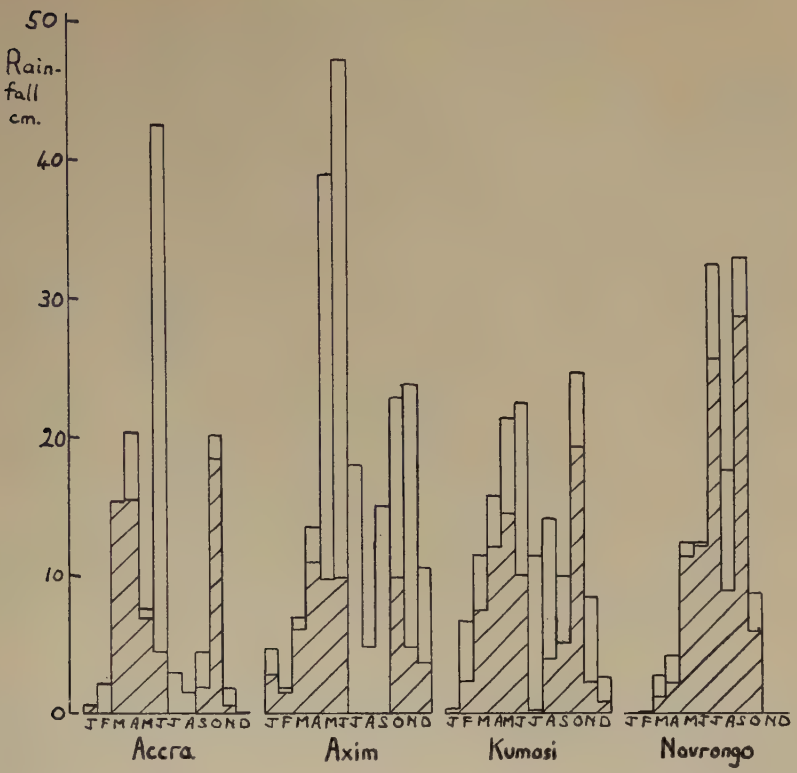


Figure 8. Proportion of monthly rainfall in 1955 due to D/Ls for Accra (5½°N 0°W), Axim (5°N 2½°W), Kumasi (7°N 2°W) and Navrongo (11°N 1°W). Amount due to D/Ls is shaded.

For three-quarters of the year almost all the rain at Accra was associated with D/Ls. This is partly owing to local topography. Large cumulo-nimbus clouds occur regularly over the scarp to the north of Accra and over an isolated outcrop to the north-east, but local showers rarely occur at Accra. Most of the rain in the other three months was not associated with thunderstorms. Some of this was undoubtedly related to the passage of D/Ls further north (see § 6) but this is not included as due to D/Ls. The ground is covered with scrub.

Kumasi lies in the forest area inland. Apparently isolated showers often occur during the D/L season.

At Navrongo D/Ls contribute a large proportion of the rainfall in all the wet months. The vegetation is of the savannah type.

Axim lies in the forest area on the coast. It shows the lowest proportion of D/L type rain of the four. Much of the rain is not associated with thunderstorms.

8. Upper-wind structure and inter-tropical front

The broad outline of the description of the upper-wind structure given by Hamilton and Archbold is generally accepted. It cannot necessarily be applied to longitudes away from those of Nigeria but it is considered by the writer to hold good over much of the area considered in this paper. Certain comments should, however, be made on this necessarily simplified picture :

(a) The ' equatorial ' or Harmattan easterly and the westerly anti-trade are by no means steady zonal currents; at 10,000 or 20,000 ft there may be winds with marked

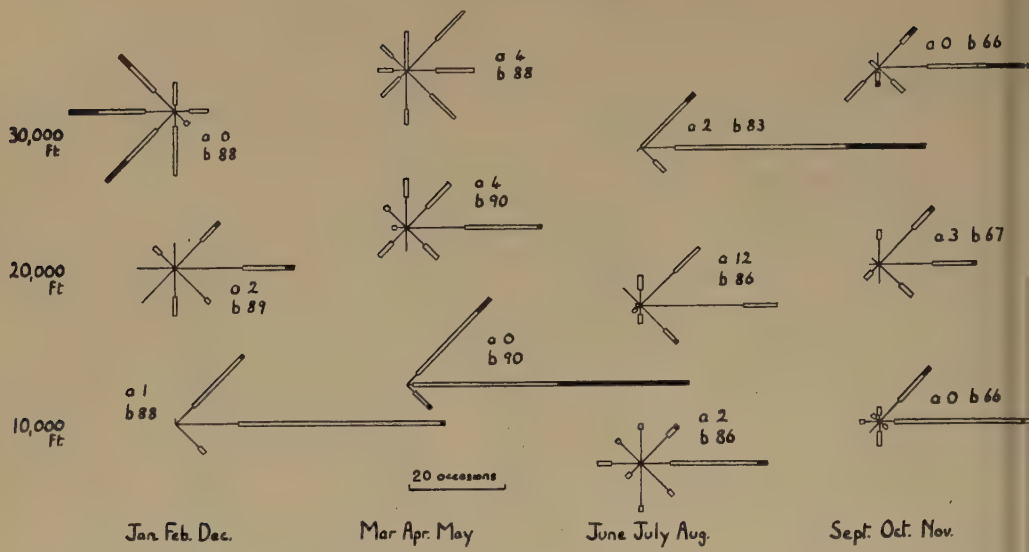


Figure 9. Upper wind roses for Lagos (Ikeja) 1955.

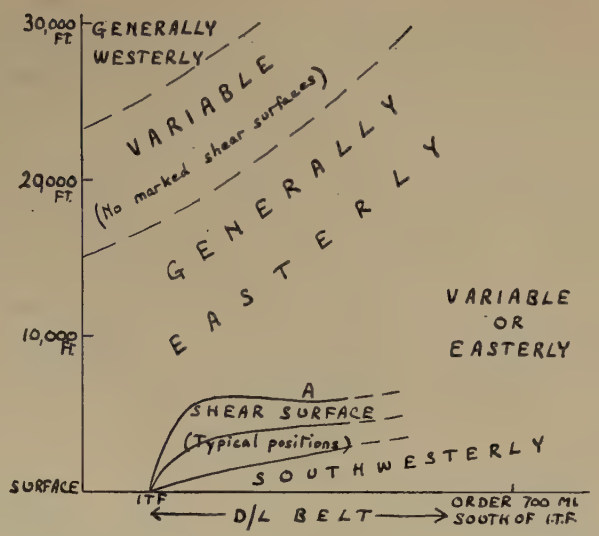
Number of occasions of wind speed 5-13 kt
,, ,, ,, ,, ,, ,, 14-27 kt
,, ,, ,, ,, ,, ,, ⩾ 28 kt

(a) indicates occasions with wind less than 5 kt. (b) indicates total number of occasions. January, Februar and December indicate dry-season conditions. March to May and September to November indicate condition in the D/L season. During June to August D/Ls are mainly to the north of Lagos.

northerly or southerly components (see Fig. 9). Over the area immediately to the south of the I.T.F. these may be significant in steering D/Ls towards a direction other than west or west-south-west as is usual.

(b) Well south of the I.T.F. (as at Lagos in June-August, Fig. 9) winds at 10,000 f are frequently light and variable rather than definitely easterly.

(c) The boundary between the dry Harmattan air-stream and the moist, south westerly air-stream appears on the surface as a fairly sharp discontinuity. It is this moisture discontinuity which is referred to in this paper as the Inter-tropical Front (not necessarily the position of the surface trough). It approximates to the line of the 60°F dew point South of this the dew point rises over a relatively short distance (order 100-200 mi) to 70°F or over. North of it, the dew point falls rapidly, not infrequently by 40°F in 100 mi or at an even greater rate. Above the ground for a few hundred miles south of the I.T.F the change from southwesterly to easterly winds coincides with the change from relatively moist to relatively dry air but, as pointed out by Hamilton and Archbold, further south the moisture discontinuity becomes less marked. The height where the south-westerlies change to easterlies is generally assumed to be the height of the ' frontal ' surface but while this point frequently indicates a fairly steady increase in height from north to south it does not always do so. For example, on 7 February 1955 the I.T.F., orientated approximately east-west, was south of Navrongo (11°N 1°W) but at Tamale, 90 mi farther south, marked south-westerlies existed to 5,000 ft indicating a steeply sloping 'frontal' surface. A statistical analysis of upper winds at Accra (6°N 0°W) and Tamale (9°N 1°W) nearly 300 mi farther north shows that when the I.T.F. is more than about 200-300 mi north of Ghana there may be little or no slope to the 'frontal' surface or it may even be higher in the north than in the south (see A, Fig. 10). Again, the transition from monsoon to Harmattan air may be through a layer of light variable, southerly or northerly winds.



igure 10. North-South cross-section through the atmosphere in relation to the I.T.F. showing wind directions over West Africa for longitudes approximately 10°E-5°W.

he above factors combine to give the structure suggested in Fig. 10 which is somewhat ifferent from that put forward by Hamilton and Archbold.

(d) West of about 10°W in the middle months of the year there is usually no sharp ew point discontinuity between the air-stream with southerly and northerly components nd the I.T.F. is not clearly marked.

The area affected by D/Ls lies south of the I.T.F. and moves north and south with ne seasonal movement of this feature. The average latitude of the I.T.F. over or near Ghana in 1955 is shown in Table 3. At 15°E it is a degree or two further south and at 5°W it is a degree or two farther north. In any month the I.T.F. may be a few degrees ither side of the positions quoted.

TABLE 3. AVERAGE LATITUDE OF I.T.F. OVER OR NEAR GHANA IN 1955

Month	J	F	M	A	M	J	J	A	S	O	N	D
osition of I.T.F. eg. N. of Equator)	8	9	10½	13½	14	17	19	19	18	14	10½	7½

9. NATURE OF THE DISTURBANCE LINE

Hamilton and Archbold stress that the nature of a Disturbance Line is similar to nat of a local thunderstorm. They also suggest a theory to account for the speed f D/Ls in terms of the speed and properties of the main air streams (p. 259). It is suggested ere that the general features of D/Ls so far established can all be related to the mechanism f the local thunderstorm. This seems preferable in view of the fact that the passage f a D/L generally increases the westward momentum of the air, at least in the lower alf of the atmosphere. Were the propagation to be accounted for mainly by the existing ir streams, one would expect the westward momentum to be decreased.

Sawyer (1949) has shown how thunderstorms may move away from high ground and there is little doubt that this explanation accounts for the *start* of many D/Ls. While the orientation of the high ground does account for the orientation of some D/Ls (notably those parallel to the coast from Liberia to French Guinea) further explanation is needed for the growth and orientation of many others. In the same paper Sawyer draws attention to the region of high surface pressure which frequently forms under a precipitating cumulo-nimbus cloud. In the D/L belt south of the I.T.F. in West Africa an isolated cumulo-nimbus cloud tends to drift in a westerly direction. The surface pressure will therefore be higher under the cloud than in areas surrounding the cloud with the exception of the area to the rear of the cloud, i.e., to the east. Thus the squall will tend to blow in all directions except to the east. This is shown by Fig. 3 of Hamilton and Archbold's paper where the squalls included are from both D/Ls and local thunderstorms. The squall will therefore assist in the development of the cloud on all sides except the east. A north-south line of clouds will thus begin to form and it will tend to grow on the west side and dissipate on the east. We thus have the beginning of a movement through the air-mass with the subsidence mainly in the rear of the D/L. It may be noted that the rate of advance of the D/L through the air at 10,000 ft (§ 5) is of the same order as the rate at which a D/L increases in length in its early stages (§ 4).

As the Coriolis factor is relatively small in the latitudes under consideration it is probable that the increase in upper wind speed behind many D/Ls (§ 5) is associated with the east-west pressure gradient at upper levels related to the cooling of the lower layers of the air column in the rain area and the subsidence warming of the air in the lower part of the atmosphere in the rear of the D/L. The accumulation of air due to the increase of wind speed with substantially the same direction will be disposed of, at least partly, by the subsidence of the air, but it may also be disposed of in an upward direction. This will account both for the persistence of medium cloud for several hundreds of miles in the rear of some D/Ls and for the observed secondary effects.

Acknowledgments

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Integrated absorbtion by the 9·6 μ band of ozone

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Summary

An account is given of laboratory measurements of absorption by the 9·6 μ band of ozone using dry air as the foreign gas. The experimental results for the band area, A_τ, are tabulated for ozone amounts up to 1·5 cm STP and pressures from 10 to 760 mm Hg at a temperature of 20°C; the slit width is 6·5 cm^{-1}. An empirical formula for A_τ is given. The best fit is obtained with slightly modified values of the ultra-violet absorption coefficients determined by Vigroux (1953), when the standard error in band area is 2·4 per cent.

1. Introduction

For the study of atmospheric ozone by observations on the 9·6 μ rotation-vibration band in the solar spectrum and in the terrestrial emission spectrum (Walshaw and Goody 1956, Goody and Roach 1956, Epstein, Osterberg and Adel 1956), it is essential to have accurate knowledge of the dependence of band absorption on the amount of ozone in the path and the ambient air pressure. The measurements described here were made with this end in view, aiming at an accuracy of one per cent in the band area A_τ ($= \int A_\nu \, d\nu$ taken over the whole band if A_ν is the fractional absorption at frequency ν) over the range of ozone amounts m from zero up to 1·5 cm STP and the range of pressures p from 0·5 to 760 mm Hg.*

At sufficiently low pressures the variation of A_τ with m and p is dominated by the Doppler broadening of the unresolved absorption lines; a discussion of the results in this region, which sets in at about 10 mm Hg, has been published (Walshaw 1955). Above this pressure, collision broadening predominates, and this is the region of most importance for the interpretation of atmospheric observations made from a ground station. This work has so far been only very briefly reported (Walshaw and Goody 1954) and the results published in a form unsuitable for general use (Fig. 2, Walshaw and Goody 1956).

2. Experimental

Pure ozone was prepared by fractional distillation of ozonised oxygen coming from a silent discharge tube (Vigroux 1949). It was introduced into an evacuated cross-shaped absorption tube and its infra-red spectrum recorded through one arm after the admission of successive doses of dry air (see Fig. 1). All the spectra were taken at the same slit-width, equivalent to 6·5 cm^{-1} at the centre of the band. Before and after each spectrum a spot reading was taken of the ultra-violet absorption in order to determine the amount of ozone in the tube; for this purpose light from a mercury lamp traversed the other arm. Absence of pressure effects in the ultra-violet region has been assumed from previous careful investigations (Strong 1941, Barbier and Chalonge 1942).

The total pressure was measured during the scanning of the infra-red spectrum. The amounts of air admitted were usually such as to double the existing pressure, until the attainment of atmospheric pressure terminated the run.

* Earlier unpublished measurements by Summerfield (1941) have been employed by some workers, but are insufficient to cover adequately the required range of m and p. In addition m was measured by a somewhat indirect method which could easily give rise to systematic errors in the results.

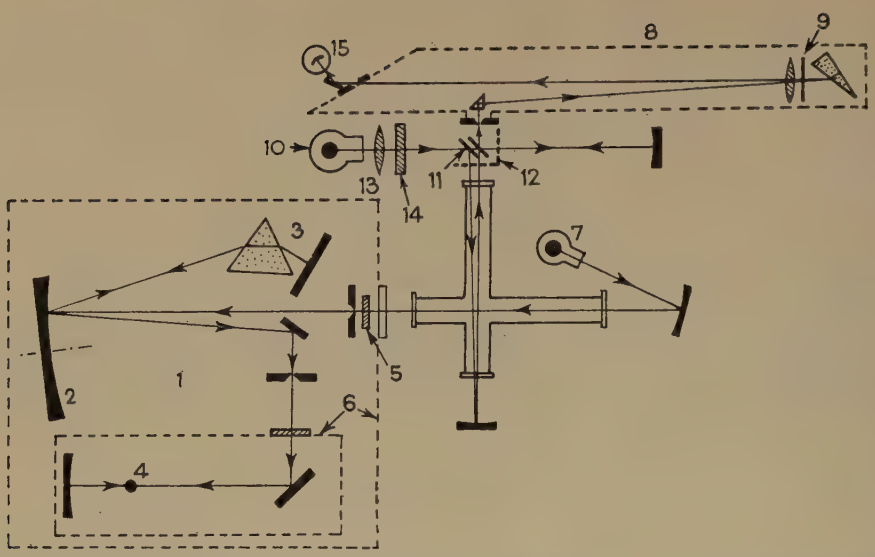

Fig. 1. Diagram of apparatus (not to scale). The *infra-red spectrometer* (1) is of conventional design with off-axis collimating mirror (2), 60° rocksalt prism with Littrow mirror (3), and thermocouple detector (4). A perspex shutter (5) operated electromagnetically is used to record the zero level. A large amount of thermal insulation (6) is provided. Light from the Nernst filament (7) is focused on the entrance slit after passing through the absorption tube. The *ultra-violet spectrometer* (8) has a shutter at (9) to assist the elimination of stray light. The beam from the mercury lamp (10) is divided by the two small mirrors (11); one part passes twice through the absorption tube and falls on the bottom half of the spectrometer entrance slit, while the other, falling on the top half of the slit, acts as reference. One or other beam is selected by a shutter (12). A lens (13) forms an image of (10) inside the absorption tube and is itself imaged on the slit, an arrangement which reduces sensitivity to slight movements. The exit slit and photomultiplier are at (15), (14) is the filter of Chance OX7 glass.

Before and after each run the tube was evacuated and flushed out with air to remove traces of ozone; infra-red 'background' spectra were then taken together with ultra-violet readings for zero absorption.

The infra-red technique was based on that of Goody and Wormell (1951). The spectrometer was of conventional optical design using a 60° rocksalt prism of 7 cm base in a modified Littrow mounting. The detector was a Hilger-Schwarz d.c. vacuum thermocouple whose output was amplified by a galvanometer amplifier and recorded photographically. By careful thermal insulation of the spectrometer and in particular of the thermocouple chamber and galvanometer amplifier parts, and by control of the room temperature (20°C) by a thermostat, effectively complete freedom from drift was achieved. The Nernst filament source could also be made to operate very steadily by carefully shielding it from draughts; its current was stablilized by barretter tubes.

By reducing zero drift and source fluctuation the random errors were reduced well below the level of reading accuracy of the photographic trace. Systematic errors in the infra-red absorption can arise from (a) stray light of shorter wavelengths reaching the detector (b) non-linearity of the amplifier system (c) convergence of the light through the absorption tube towards the source image on the entrance slit (d) errors in measurement of tube length, and (e) impurities in the ozone. These points were all carefully investigated (Walshaw 1954) and the total systematic error in A_r is believed to be less than one per cent.

The ultra-violet absorption measurements were made with a photomultiplier (type 27M3) behind a slit mounted in the focal plane of a Hilger E.1. quartz spectrograph. A double-beam optical system was used in order to avoid errors caused by the variation of sensitivity over the cathode surface. The photomultiplier output was measured with a

potentiometer circuit using a valve-voltmeter to indicate balance. It was assumed that the photomultiplier response was strictly linear at the light intensities used (see, e.g., Engstrom 1947). Stray light was reduced by a filter of Chance OX7 glass and careful choice of shutter position; the remainder was measured and corrected for, although the correction was usually very small since one could choose a mercury line whose absorption was not too strong.

The stability of the spectroscopic apparatus, which was very sturdily supported on a concrete bench in a cellar, was evident from the almost perfect agreement between the background spectra and ultra-violet measurements before and after runs. The very few runs showing poor agreement were rejected.

The pressure was measured on a mercury barometer and a two-stage MacLeod gauge designed to cover the whole pressure range with an accuracy of 0·5 per cent or better. To prevent contact between ozone and mercury, a sensitive glass spoon-gauge was employed as a null indicator between the absorption tube and the air-filled barometer system.

The tube length was 34·9 cm in the infra-red beam and 37·2 cm (traversed twice) in the ultra-violet beam. In addition a number of runs were made with a shorter absorption tube (6·5 cm) to measure the different broadening effects of ozone and air molecules (as discussed by Goody and Wormell 1951). The conclusion, that an ozone molecule is 1·61 times as effective in collision-broadening as an air molecule, has been used to correct for the large partial pressure of ozone in the absorption tube and so to reduce the results to the atmospheric condition of infinite dilution in air.

3. Results

Measurement of the infra-red spectra enabled curves of absorption against arbitrary units of prism rotation. (' flash numbers ') to be plotted.* The area under these curves was measured by planimeter, separating the 9·6 μ from the much weaker overlapping 9·0 μ band by a consistently applied graphical procedure. The band areas A_τ were then converted to cm^{-1} units with the aid of the wavelength calibration of the spectrometer. The results are given in Table 1.

For various reasons it is desirable to represent measurements of this kind by a suitable formula. The problem of smoothing a function of two variables is particularly severe when only a limited number of observations is available, and is even worse when, as in atmospheric applications of the present data, the first derivative of the data is the really important quantity. In addition a formula is more concise than a table, facilitates a quantitative assessment of experimental error, and lends itself readily to electronic computation.

It has been found that A_τ can be accurately represented by an expression of the form

$$A_\tau = \Delta\nu\,[1 - 10^{-\mu f(\phi)}] \quad . \quad . \quad . \quad . \quad (1)$$

where $\Delta\nu$ is the frequency range over which the band absorbs, $\mu = m\zeta$ and $\phi = m\zeta^a/p$; a is a constant and $\zeta(m)$ an arbitrary function of m only. In the special case when $\zeta = a = 1$ and $f(\phi) = A(1 + B\phi)^{-\frac{1}{2}}$, Eq. (1) reduces to the absorption by a simplified band model discussed by Goody (1952); here, however, the more general form is intended to have no theoretical significance whatever, serving only as an empirical formula. It has the practical advantage that it may readily be fitted to the data by graphical methods which the interested reader may easily derive as a slight extension of that indicated by Walshaw and Goody (1954).

* As it is impracticable to publish the complete spectral data, a microfilm is being prepared which will be available on loan from the library of the Royal Meteorological Society.

TABLE 1.

m = ozone amount in cm STP assuming the absorption coefficients given under (2) in Table 2; p = pressure in mm mercury corrected to infinite dilution; A_τ = band area in wave numbers; ϵ = excess of observed over calculated band area, as percentage of observed. (The symbols in the first column refer to Fig. 2).

Run λ	$10^3 m$	p	A_τ	ϵ	Run λ	$10^3 m$	p	A_τ	ϵ
	26·0	11·2	5.70	+ 4.8		481	17·6	35·6	0
	26·8	25·1	7·09	+ 3·1		503	20·8	38·8	+ 1·3
a	27·2	47·9	7·83	+ 3·3		503	29·0	42·4	+ 1·1
2894	27·6	99·0	8·57	+ 3·3	i	500	40·6	45·5	0
⊙	27·7	216	8·82	+ 0·7	3132	501	58·5	49·1	− 0·2
	27·8	449	9·01	0	+	497	85·4	52·6	− 0·9
	28·0	729	9·08	0		486	165	57·8	− 1·5
						468	328	62·5	− 0·9
	38·2	13·9	8·02	− 3·5		425	734	65·8	+ 0·6
	39·2	26·5	9·55	− 4·0					
b	38·9	53·5	10·82	+ 1·1					
3021	39·6	104	11·92	− 0·7		594	43·0	50·2	+ 2·3
○	39·5	217	12·67	+ 0·4	j	604	89·7	58·0	+ 2·0
	39·1	418	13·14	+ 2·2	3132	601	166	62·7	+ 0·5
	38·9	727	13·30	+ 2·9	▲	591	330	65·3	− 1·9
						559	735	68·9	− 0·6
	63·7	11·5	11·47	+ 9·0					
	68·4	24·4	13·46	− 0·2					
c	69·4	42·1	15·66	+ 1·1		686	24·9	46·0	+ 1·4
3021	70·0	80·6	17·94	+ 1·5		724	29·6	49·0	+ 2·2
△	70·5	227	20·3	+ 0·6		724	38·9	52·0	+ 1·6
	65·6	655	21·1	+ 4·8	k	724	53·4	55·6	+ 2·5
					3132	724	78·2	58·4	0
	102	24·9	17·86	+ 1·3	■	719	151	64·2	− 0·5
	101	32·2	19·09	+ 2·5		698	290	68·6	+ 0·3
d	100	65·5	22·5	+ 2·3		651	578	70·7	+ 0·2
3021	99·9	130	24·6	+ 1·0		633	739	71·3	+ 0·5
□	99·5	258	26·9	+ 1·7					
	95·3	496	28·2	+ 5·2					
	94·6	744	28·5	+ 3·4		1036	76·6	64·7	+ 0·2
					l	1078	113	69·1	+ 0·5
	158	12·7	18·11	− 1·5	3342	1050	214	73·0	− 0·5
	162	23·0	22·0	− 2·1	◆	1050	453	75·6	− 1·5
e	163	42·9	26·0	− 2·5		1004	743	76·7	+ 0·5
3021	166	82·3	30·4	− 2·6					
▽	167	166	34·9	− 1·0					
	166	329	38·0	− 0·2		1210	42·9	62·1	+ 2·0
	162	654	39·3	− 0·9		1316	48·6	64·9	+ 1·5
					m	1313	68·1	67·6	− 0·2
	196	26·7	26·8	+ 2·5	3342	1313	96·0	71·3	− 1·6
f	201	45·2	29·5	− 3·4	▼	1284	174	75·6	+ 0·8
3021	201	82·9	34·2	− 2·0		1256	334	78·0	− 0·4
◊	201	168	38·6	− 1·8		1187	743	79·4	− 1·2
	196	744	44·7	0					
	214	13·2	21·3	− 2·1	n	1968	66·9	74·1	− 0·9
	221	32·2	29·4	− 0·2	3342	1941	293	83·5	− 1·7
g	223	72·0	34·1	− 4·8	●	1861	538	84·2	− 2·1
3021	226	141	39·5	− 3·6		1795	753	84·2	− 2·1
♂	224	288	42·9	− 4·0					
	215	731	46·0	− 2·7	o	5·58	728	2·00	− 2·8
					2536	8·09	728	2·97	+ 0·2
	365	14·7	30·2	+ 2·0	Ø	8·76	728	3·28	+ 2·1
	370	18·9	32·6	+ 0·4					
	371	25·4	35·3	− 0·6		2·78	728	1·01	+ 3·9
	371	40·0	38·5	− 4·6		6·35	728	2·13	− 3·2
h	375	50·2	41·7	− 1·9		7·55	728	2·52	− 3·7
3132	370	74·8	45·1	− 2·3	p	10·8	728	3·83	+ 3·7
×	367	151	50·5	− 3·8	2894	17·0	728	5·63	− 1·8
	356	287	55·1	− 2·2	Ø	17·6	728	5·99	+ 1·6
	331	564	58·9	+ 0·5		21·3	728	7·11	+ 0·2
	324	736	59·1	+ 1·7		24·0	728	7·98	+ 0·8

On fitting this formula to the results it was found that the deviations varied systematically with the wavelength used for the measurement of m. A reduction in these trends could be effected by modifying the absorption coefficients used to convert the measured ultra-violet absorption to ozone amount. The original values, obtained from the published data of Vigroux (1953) are given in Table 2 together with the modified values. It will be noted that no change has been made for the lines 3021 and 3132 A, which account for 5 per cent of the observations. The change at 3342 A is hardly significant because it is very difficult to interpolate from Vigroux's figures in this region. At 2536 and 2894 A, however, the marked band structure has disappeared and interpolation is less uncertain; nevertheless the modified values give significantly greater consistency to the present data and have consequently been adopted for the ozone amounts listed in Table 1.

The confirmation of Vigroux's value for the absorption coefficient at 3021 A by an independent method is reported in this issue by Hearn, Walshaw and Wormell (1957). In addition Mr. Hearn has kindly allowed me to quote his recent preliminary measurements on 3342 and 2894 A: his values are 0·049 and 17·2 respectively, supporting the modified values given in Table 2.

TABLE 2.

λ	No. of spectra	Absorption coefficient, base 10 (cm STP)$^{-1}$ Vigroux (1)	Adopted (2)	$\bar{\epsilon}$
2536	3	124	115·7	− 0·17
2894	15	18·2	16·0	+ 1·11
3021	38	3·29	3·29	− 0.01
3132	33	0·76	0·76	+ 0·01
3342	16	0·070	0·0504	− 0·43

The fitting of Eq. (1) to the data was done graphically, as mentioned above, yielding the results :

$$\Delta\nu = 138 \text{ cm}^{-1} \qquad a = 2{\cdot}11$$

$$\zeta(m) = \begin{cases} \dfrac{1 + 0{\cdot}1025\, m}{1 + 1{\cdot}61\, m} & \text{for } m \leqslant 0{\cdot}1 \\ 0{\cdot}984 \times 10^{-0{\cdot}53\, m} & \text{for } 0{\cdot}1 \leqslant m \leqslant 0{\cdot}4 \\ 0.317 \times m^{-0{\cdot}74} & \text{for } m \geqslant 0{\cdot}4 \end{cases}$$

$$f(\phi) = 1{\cdot}185\,(1 + 734\,\phi)^{-\frac{1}{2}}\,\eta(\phi)$$

u is in cm STP and p is in mm Hg. The function $\eta(\phi)$ is tabulated in Table 3.

TABLE 3.

$\log_{10}\phi$	$\eta(\phi)$	$\log_{10}\phi$	$\eta(\phi)$
$\leqslant -4{\cdot}0$	1	− 3·0	0·977
− 3·9	0·998	− 2·9	0·979
− 3·8	0·995	− 2·8	0·982
− 3·7	0·993	− 2·7	0·989
− 3·6	0·989	− 2·6	1·002
− 3·5	0·984	− 2·5	1·016
− 3·4	0·982	− 2·4	1·030
− 3·3	0·980	− 2·3	1·045
− 3·2	0·977	− 2·2	1·057
− 3·1	0·977	− 2·1	1·069
		− 2·0	1·079

In order to test the accuracy of the fit and to estimate the experimental error, the excess of observed over calculated band area has been computed for each point. This quantity, expressed as a percentage of the observed band area, is called ϵ and is included in Table 1 and plotted in Fig. 2. The use of ϵ as a criterion of fit rather than another quantity (e.g., difference in transmission) leads to an estimate of error weighted as far as possible in the same way as the experimental error. A constant fractional error in the measured quantity also accords with most practical requirements for data of this type.

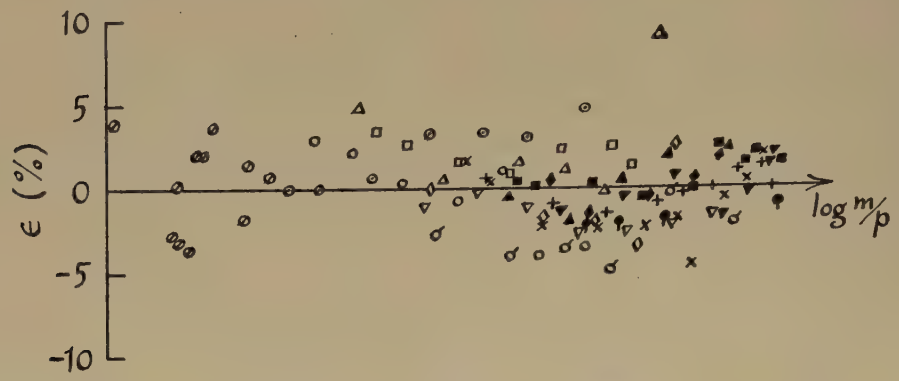

Fig. 2. Comparison of observed band area with values calculated from the empirical formula. For symbols see Table 1.

Careful examination of Fig. 2 shows the residual systematic error in each run to be very small. When the points are grouped according to ultra-violet wavelength, again no systematic trends are detected (Table 2); it therefore appears that Eq. (1) represents the data quite adequately. The mean value of ϵ is + 0·09, while the r.m.s. value is 2·4. The latter figure may be taken as the overall random error (per cent) of the measured band area, and is satisfactorily close to the error expected from the combination of the various errors of measurement.

It should be pointed out that earlier reductions of these measurements employed the ultra-violet absorption coefficients of Ny and Choong (1933). With these coefficients the systematic differences between the different wavelengths are considerably bigger than with the unmodified Vigroux coefficients. Their effect on the choice of constants in Eq. (1) is fortunately not great enough to have a significant influence on the results of Walshaw and Goody (1956). Since the measurements of Hearn, Walshaw and Wormell (1957) now decide unequivocally in favour of Vigroux's coefficients, the formulae given in this paper should be used for future applications of the 9·6 μ data. It must be borne in mind, however, that the international network of stations using Dobson spectrophotometers at present employs the Ny and Choong coefficients; its ozone values must therefore be multiplied by a factor approximately equal to 1·30 to make them consistent with these data. The exact value of this factor is at present under consideration by the Ozone Commission in preparation for a change to the Vigroux scale.

4. Acknowledgments

The experimental work was done at the Cavendish Laboratory and my sincere thanks are due to Dr. T. W. Wormell and Dr. R. M. Goody for their interest and encouragement. I am also indebted to the D.S.I.R. for a grant held at this time.

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* Through the courtesy of Dr. John Strong a microfilm copy of this document is now in the library of the Royal Meteorological Society.

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Dew

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SUMMARY

Dew formation on short grass has been studied with a balance, recording weight of condensation, and with filter papers to absorb moisture. Three regimes are distinguished: in the first, daytime evaporation continues and the grass remains dry; in the second, the surface continues to lose weight but the grass becomes wet owing to the partial condensation of water vapour evaporating from the soil; in the third, this loss of weight ceases or there is an increase in weight attributed to condensation of atmospheric water vapour. A distinction is therefore made between ' distillation ' of water vapour from soil to grass (1-2 mg cm^{-2} hr^{-1}) occurring on very calm nights through a laminar layer with a transfer coefficient approaching the molecular value; and ' dewfall,' the turbulent transfer of water vapour from the atmosphere, negligible when the wind at 2 m falls below 0·5 m/sec but reaching 3-4 mg cm^{-2} hr^{-1} with stronger winds. This distinction is supported by analysis of the surface heat-budget. On windless nights, since heat flux from the soil and net radiative loss were almost equal, the transfers of sensible and latent heat from the atmosphere were negligible and distillation was of much greater importance than dewfall. Implications for taller crops and warmer climates are briefly considered.

1. INTRODUCTION

In his classic *Essay on Dew*, Wells (1815) explained his many experimental results in terms of the condensation of water vapour from the atmosphere. Dew formed most readily on calm clear nights yet ' a slight agitation of the air . . . when the air is pregnant with moisture . . . will render greater the quantity of dew.' Wells rejected previous ideas of upward movement from the soil but Aitken (1885) revived them on the basis of experiments with shallow trays of turf and bare soil from which there was a *loss* of weight on occasions when dew was forming. Aitken concluded that evaporation was taking place owing to diffusion of moisture to the surface from warmer soil beneath. He made a useful distinction between dewdrops and other waterdrops sometimes seen on leaves produced by exudation under the action of root pressure.

After Wells and Aitken, there was a long period that added little to fundamental knowledge of the physics of dew formation. As there are now various meteorological techniques available for the study of the flux of water vapour near the ground, the dew problem is worth re-examination as a special case.

Treating water-vapour flux as a turbulent transport problem, Thornthwaite and Holzman (1942) measured wind velocity, temperature, and humidity up to 8 m over a meadow, and used an integrated form of the one-dimensional transfer equation to compute mean hourly values of the flux at the surface. During the night absolute humidity generally increased with height, indicating a downward flow of vapour associated with dew formation. Working over a smaller height range and with more sensitive apparatus, Pasquill (1949a) used the same equation to calculate the diurnal variation of evaporation from clayland pasture in May. Again, the humidity gradient indicated dew formation at night, the maximum computed rate being an order of magnitude less than the maximum evaporation rate. It has therefore been assumed by many meteorologists that the condensation of atmospheric water vapour brought to the surface by turbulent diffusion is a major source of dew.

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Water-vapour exchange at the earth's surface has also been investigated in terms of e energy balance. The work of Fransilla (1936) and of Rider and Robinson (1951) ggests that an appreciable fraction of the radiative heat loss at night may be compen- ted by latent heat of condensation. On the theoretical side, Frost (1948) assumed that uring the night, heat conducted upwards through the soil is negligible compared with eat brought to the surface by turbulent diffusion and showed that the heat released in ew formation could diminish by several degrees the total nocturnal cooling.

2. Apparatus and site

With very few exceptions, instruments previously used for measuring dew have mployed an artificial collecting surface, e.g., gypsum, metal, or painted wooden plates, ut at best such devices can provide only a very rough indication of how the condensation n natural surfaces might vary from hour to hour or from night to night. The successful neasurement of condensation, like the direct measurement of surface friction, demands a est surface which resembles its surroundings as closely as instrumental design will allow. Craddock (1951) described a balance to be sunk in the ground, which gave a continuous ecord of dewfall on a natural soil or plant sample and an instrument of similar type was uilt for the present work (Jennings and Monteith 1954). This gave a photographic ecord of the change in weight of a block of turf and soil, 330 cm^2 in area and 25 cm eep, which had been carefully separated from its surroundings without destroying the oil structure.

The sensitivity was approximately 2 mg/cm^2 per cm of record and a change in weight ould be estimated to about 0·2 mg/cm^2.

A more direct, if less accurate, measurement of condensation provided valuable nformation complementary to that supplied by the balance. On several nights the noisture on the grass round the experimental site was sampled every 2 hr by absorption vith filter papers of area 400 cm^2. These were pressed down on the leaves, rotated to absorb s much moisture as possible, and immediately weighed against a piece of dry filter paper. Five papers were used for each observation to reduce sampling error, and the linear ncrease of moisture with time, a striking feature of the observations, suggested that the nethod was giving at least a good relative estimate of rates of dew formation. It was mpossible to obtain accurate absolute measurements in this way and it is believed that igures given by filter-paper absorption with this type of plant cover may underestimate ondensation by 50-100 per cent.

Temperature and humidity measurements were made with differential wet-and- lry-bulb thermocouple circuits (36 s.w.g. copper-constantan) and circuits measuring dry- ulb temperature with the 'cold' junction held in a vacuum flask of crushed ice (26 .w.g. copper-constantan). The junctions were freely exposed to the air on a Tufnol ramework with 2 cm between the junctions and the points of support. Simple aluminium adiation shields were used but since the observations were to be made almost entirely in he absence of sunshine and often in very calm conditions, it was decided that aspiration vould be both unnecessary and undesirable. A wind-tunnel test showed that the wet bulb vhich was wrapped with cotton thread for 2 cm on either side of the junction and fed rom a reservoir at both ends, gave a minimum reading for wind speeds below the stopping peed of a Sheppard cup anemometer (*c.* 30 cm/sec). 6 of these thermocouple units were xposed at heights of 3·5, 12·5, 25, 50, 100 and 200 cm. An additional set of junctions ave wet- and dry-bulb temperatures within the grass cover at 1 cm above the soil surface nd a junction pressed into the ground and covered with a very thin layer of soil gave an pproximation to the soil-surface temperature.

While the use of relatively small elements made it possible to dispense with aspiration the response time was such that spot readings of thermocouple output would often have had little significance. It was therefore necessary to take 10 readings from each element over a period of 10 min., later averaging to obtain mean profiles for the period. The galvanometer circuits had a response time of 4 sec and a sensitivity of 1°C per cm scale division, so that it was possible to express instantaneous readings to 0·1°C and mean temperatures to 0·01°C where the last figure had statistical significance only. The standard deviation of the 10 readings was generally of the order of 0·02°C.

Sheppard cup anemometers were used at 25, 50, 100 and 200 cm, the clockwork and electrical contact types being available at different times. The site of the observations, the Imperial College Sports Ground, Harlington, Middlesex, was chosen for its uniform surface and open exposure. Surrounded on its northern and western boundaries by poplar trees, the effective area of the field was about 220 × 400 m, but the disposition of the apparatus was restricted by the boundaries of cricket pitches and most of the work was carried out 25-30 m west of the observing hut which stood, 3 m high, on the east side of the field. For winds from SSW through W to NNE there was an upwind fetch of at least 150 m over level grass which was kept about 1 cm high by regular cutting. The observations of other workers (e.g. Deacon 1953) suggest that in adiabatic conditions at least, this should have been adequate to give fully-developed profiles up to a height of 2 m, but in very stable conditions when turbulent mixing was slight, the unusual humidity profiles frequently observed and occasional irregular wind profiles may have been due, in part, to lack of adequate fetch.

The dew balance sunk in its pit, and a soil heat-flux plate were the only permanent installations on the field. The remainder of the apparatus and the necessary leads had to be set up afresh before each night's work. Whenever possible, the equipment was ready for use shortly before sunset.

3. Dew and relative humidity

The variation of temperature, dew-point and relative humidity at 1 cm on three occasions of dew is shown in Figs. 1, 2 and 3, each point representing the mean of 10 observations. Of special interest are conditions at the time when moisture first appeared on the grass but for a reason later explained, it was not possible to determine this time accurately from the dew-balance trace. Times indicated by the arrows are those at which the moisture was first detected by lightly brushing the leaves with the back of the hand, a subjective, but a very sensitive test. Estimated times of dew formation may therefore be as much as a $\frac{1}{4}$ hr late since this was the period between successive examinations of the grass. It is clear that on the nights illustrated, condensation began before saturation was reached in the grass cover, and in the period September-October 1953, observed relative humidities at the time of dew formation on 7 nights ranged from 91-99 per cent., with a mean value of 95 per cent. This implies that the leaves were colder than the ambient air, and assuming that condensation began when the temperature of the leaves reached the dew point of the air, the range of air-leaf temperature differences corresponding to the observed critical humidities for condensation was 0·4-1·4°C with an average of 0·9°C. The leaf-temperature depression can be related theoretically to radiative loss, wind speed, and to the shape and size of the leaf, assuming laminar flow, but while estimates of the correct magnitude were obtained, the present observations were not sufficiently accurate to justify direct comparison between theory and experiment. One theoretical prediction was borne out qualitatively. The vegetation contained a small proportion of relatively broad-leaved species such as plantain and clover which were

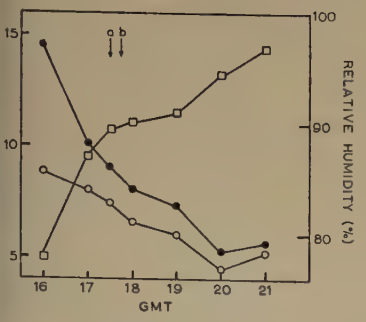

Figure 1

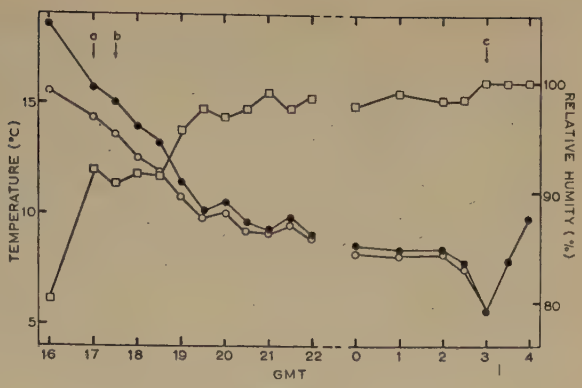

Figure 2

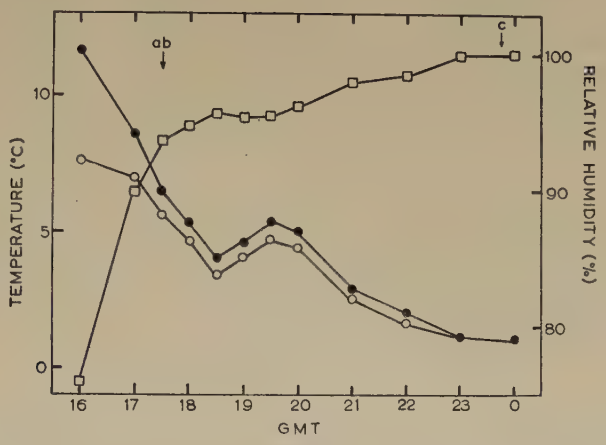

Figure 3

Figures 1-3. Variation of air temperature (●), dew-point (○), and relative humidity (□) at 1 cm in short grass cover. Arrows give time of condensation on broad-leaved species (a), on grass (b), and time when fog was first observed (c). Fig. 1: 28 Sept. 1953; Fig. 2: 1-2 October 1953; Fig 3: 5 Oct. 1953.

lways observed to become moist before the grass, presumably because their greater size vas associated with a smaller heat-transfer coefficient and hence a greater leaf-air emperature difference. It was also found that the bulb of a grass minimum thermometer xposed in the standard fashion became dewed shortly before the grass.

After condensation had begun, a check was often observed in the rate of increase of elative humidity, an effect particularly marked in Fig. 2. It appears that the onset of dew ormation was followed by a period in which water vapour was removed by the cover more apidly than it could be replaced by diffusion. But while dew formation was often ssociated with a discontinuity in the increase of relative humidity, a simultaneous emperature discontinuity was never observed either from half-hourly values of emperature or from a thermograph exposed with its element 15 cm above the ground. Discontinuities of the type described by Saunders (1949) were marked on many clear, calm ights generally some time after dew had begun to form, e.g., at 1930 GMT on 8 September (Fig. 2) and 1830 GMT on 1 October (Fig. 3). Since these temperature ncreases were accompanied by an increase in the dew point, as reported also by Pedlow 1952), they cannot have been caused by a sudden increase in the rate of dew formation ut should probably be attributed to an increase in the turbulent transfer both of sensible nd of latent heat towards the ground from warmer and moister levels above.

Later in the night, air within the cover approached saturation slowly and on a few occasions 100 per cent humidity was reached before dawn or before the development of cloud cover inhibited further cooling. Saturation was always followed by the formation of fog and Fig. 2 shows an interesting example. After 0200 GMT there was a slight drop in the wind speed, a fall in temperature, and a rise in the relative humidity within the cover from 98 per cent to saturation. At 0230 GMT, isolated patches of fog were observed in the neighbourhood but not in the immediate vicinity of the apparatus. Between 0230 and 0300 GMT cooling was intense, perhaps caused by radiation from fog droplets as suggested by Stewart (1955) and at the same time the vapour pressure fell by 1·3 mb, equivalent to a change in liquid-water content of 0·1 g/m^3. By 0330 GMT fog completely enveloped the field and with the consequent reduction in the radiative loss from the surface, temperature rose rapidly. The increase of vapour pressure at the same time suggests that evaporation of fog droplets may have taken place from the lower layers of the fog to maintain saturation. Thereafter the rapid disappearance of the fog may have been associated with the development of 8/8 stratus at 1,000 ft, which became clearly visible at 0400 GMT. Fog was never observed at the site before saturation occurred in the grass cover, and it is clear that the initiation of condensation from unsaturated air followed by a slow increase of relative humidity may be important factors in delaying the appearance of ' radiation ' fog.

4. Three regimes in water-vapour transfer

Comparison of dew-balance traces with observations and direct measurements of moisture on the grass made it possible to distinguish three regimes in the diurnal exchange of water between the atmosphere, the soil, and the grass.

(*a*) *Evaporation regime*

The evaporation regime, marked (i) in Figs. 4 to 8 is the name given to periods before or after sunset when the balance pan lost weight and the grass remained dry. The rate of loss in weight, which was of the order of 70 mg cm^{-2} hr^{-1} in full sunlight, decreased rapidly in the late afternoon. On many evenings, evaporation continued after sunset and even persisted throughout the night when cooling was slow owing to a relatively high wind or to the presence of cloud (Fig. 4). Records of humidity in the cover showed that evaporation occurred after dark only when the air in the cover was unsaturated, and that the rate of evaporation decreased as saturation was approached. Possible explanations have been discussed elsewhere (Monteith 1956). The most satisfying assumes that water vapour diffuses from a relatively warm soil surface to the cooler and drier air surrounding the leaves and thence into the free atmosphere.

Knowing the evaporation E it was possible to calculate the diffusion coefficient K_v from

$$E = -K_v \frac{\partial x}{\partial z} \simeq K_v \{x(0) - x(1)\} \quad . \quad . \quad . \quad (1)$$

where the gradient $\partial x/\partial z$ is replaced by the difference in absolute humidity between the surface and 1 cm. In a soil supporting normal unwilted vegetation, the relative humidity of the soil atmosphere in the root zone is known to be in the range 99-100 per cent. Assuming that at night the air in contact with the surface was in fact saturated, the vapour pressure at the surface was found from the surface temperature and tables of saturated vapour pressure, while wet-and-dry-bulb temperatures at 1 cm gave the vapour pressure at 1 cm from the psychrometric equation. The absolute humidity x in g/m^3 corresponding to a vapour pressure e in mb was then found from the ideal gas equation which gives $x = 0{\cdot}76\,e$ at 13°C, the approximate mean temperature of the observations.

TABLE 1. DATA FOR CALCULATIONS OF WATER-VAPOUR TRANSFER COEFFICIENT IN SHORT GRASS COVER

Date	Time GMT	x (0) (g/m^3)	x (1) (g/m^3)	V (25) (cm/sec)	E (mg cm^{-2} hr^{-1})	K_v (cm^2/sec)
7 Sept. 1953	1830	13·05	9·52	106	4·6	0·36
	1900	12·10	9·23	80	3·2	0·31
	1930	11·81	9·23	125	3·1	0·35
	2000	11·38	9·15	76	2·1	0·26
	2100	11·08	9·60	107	1·5	0·28
11 Sept. 1953	1830	13·55	10·42	206	6·0	0·53
	1930	12·70	10·28	293	3·9	0·41
	2030	11·72	9·84	204	3·1	0·46
14 Sept. 1953	1800	11·42	7·77	100	3·9	0·30
	1830	10·51	7·72	133	2·7	0·27
	1900	10·44	8·70	152	2·5	0·40
	2000	10·38	8·62	150	2·5	0·40
28 Sept. 1953	1730	9·77	8·03	98	2·8	0·45
	1800	9·06	7·40	132	2·8	0·47
	1900	8·74	7·10	104	1·7	0·29
1 Oct. 1953	1800	12·10	11·02	151	2·2	0·56
	1830	11·72	10·60	132	2·2	0·54

The data given in Table 1 for relatively windy nights when the grass remained dry, show that 10 min mean values of K_v were of the same order of magnitude as the molecular value, 0·24 cm^2/sec, suggesting a close approach to laminar flow within the cover. On a night when there was negligible temperature gradient in the lowest two metres, the wind profile fitted the standard formula for rough flow

$$u = (u^*/k) \ln z/z_0 \quad . \quad . \quad . \quad . \quad . \quad (2)$$

where k, von Kármán's constant = 0·40, z_0 = 0·08 cm and u^* = 16 cm/sec when the velocity at 2 m was 3 m/sec. These values were obtained from the mean of 4 wind profiles over a 4 hr period, each profile being in turn a 10 min mean (Table 2).

TABLE 2. WIND DATA (cm/sec) FOR 11 SEPT. 1953

Height (cm)	25	50	100	200
Time (GMT)				
1730	178	206	224	242
1830	256	288	316	368
1930	293	331	362	400
2030	204	223	249	263
Mean	233	262	288	318

In the classical concept of fully-rough flow the roughness elements of the surface project into the turbulent boundary layer and an aerodynamically-rough surface has been defined as one in which the irregularities are sufficiently large for the motion to be turbulent down to the surface itself. Present observations of water-vapour diffusion suggest that fully-rough flow may be maintained over short grass cover even when molecular processes play an important part in diffusion within the cover. If this is so, the main production of shearing stress must take place in the top few millimetres of the cover, the remaining stress generated at the soil surface and in the lower parts of the cover being associated with quasi-laminar flow among the leaves.

Since the transfer coefficient for water vapour did not exceed 0·53 cm²/sec on a night when the wind at 2 m was 2-3 m/sec, the molecular value of the coefficient may be assumed appropriate for wind speeds less than the stopping speed of the anemometers (*c.* 0·3 m/sec) which were frequently encountered during the work.

(*b*) *The transition regime*

On some nights when the temperature dropped rapidly at sunset the evaporation regime was followed by a short period during which the balance continued to record a loss of weight although moisture had begun to appear on the grass ((ii), Figs. 5 and 6). The existence of such a period, referred to as the 'transition regime' made it impossible to use the balance records to decide when dew formation began. The same behaviour of turf has been reported by others e.g. by Aitken and by Lehmann and Schanderl (1942) and since measurements of temperature and visual inspection of condensation indicated that the test specimen was cooling at the same rate as its soil environment, the effect is believed to be real.

The evaporation regime has been modified, for it is obvious that when the leaves cool to the dew-point of the air, condensation will begin and the flux of water vapour from soil to atmosphere will be correspondingly reduced. Initially, the rate of condensation is restricted by the heat balance of the surface since condensation constitutes a new source of heat within the cover. Were condensation to proceed more rapidly than the equilibrium rate, sufficient latent heat would be released to raise the temperature of the leaves above the dew point, preventing condensation until further cooling had taken place. The idea that dew formation may be a self-quenching process of this kind was supported by Seeman and Löw (1944) who attributed rises in air temperature of as much as 2°C at 10 cm to the formation of dew. These workers, measuring dew as the increase in weight of a metal plate, showed that the rate of condensation on the plate fluctuated in phase with temperature fluctuations with a period of roughly 20 min. A natural grass surface has very different thermal characteristics and it cannot be assumed that condensation on grass will show the same features as that on a plate. In the present observations there was never any evidence that dew formation was periodic although temperature fluctuations were always observed on calm nights.

The initiation of dew was followed by a slow increase in the condensation rate, controlled, presumably, by a fine balance between the terms in the heat budget of the surface. Since the evaporation rate always fells to zero within about an hour of the first sign of dew, it appears that the heat balance was able to adjust itself within that time, enabling all the vapour diffusing from the soil to condense within the cover.

(*c*) *The condensation regime*

Nights on which measurable condensation occurred ((iii), Figs. 5-8) can be subdivided into two types :

(a) those nights on which the rate of condensation recorded by the balance was at least an order of magnitude less than the rate of condensation measured by filter-paper absorption of moisture on the grass. Occasionally, the weight record of the balance ran parallel to the time axis for several hours (Fig. 6), indicating that the net exchange of water vapour at the surface was less than the observational error i.e., $< 0{\cdot}2$ mg/cm²,

(b) those nights on which the weight increase of the balance and filter-paper absorption gave condensation rates of the same order of magnitude (Figs. 7, 8).

It is believed that the difference between these two types is to be attributed to wind speed; for type (a) nights were those of very light wind (0·5 m/sec or less at 2 m), clear

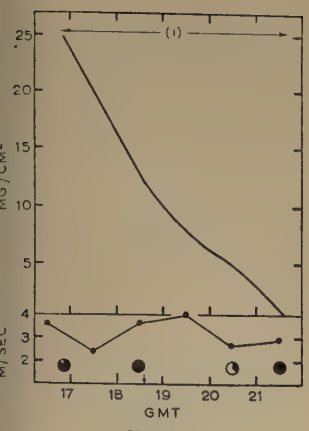

Figure 4

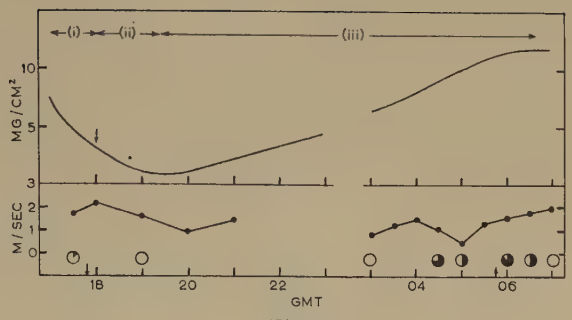

Figure 5

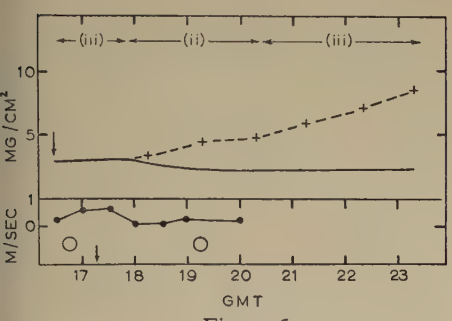

Figure 6

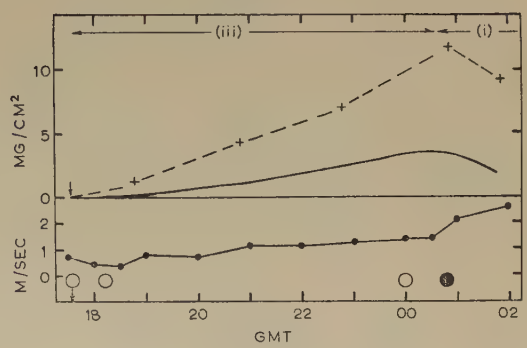

Figure 7

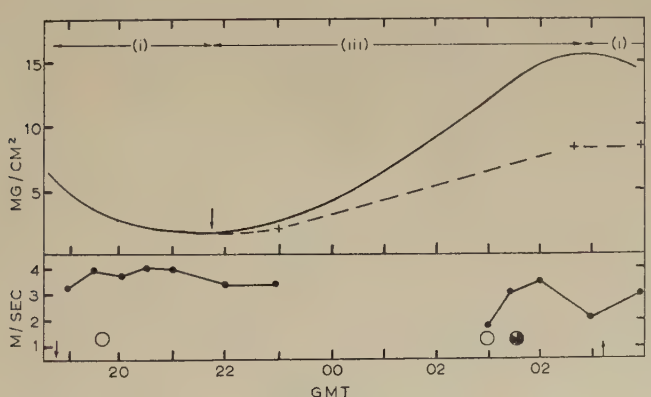

Figure 8

igure 4-8. Dew balance record (full line) and filter-paper absorption measurements (broken line) with wind peed at 2 m and cloud amount indicated by black segment of circle. Arrows against balance record ndicate observed time of condensation on grass; arrows against time axis give sunset and sunrise. Fig. 4 : 1 Sept. 1953; Fig. 5 : 1-2 Oct. 1953; Fig. 6 : 9 Oct. 1953; Fig. 7 : 3-4 Oct. 1953; Fig. 8 : 31 Aug.-1 Sept. 1953.

kies, and consequently rapid cooling, while on type (b) nights, although cooling was enerally sufficiently rapid for condensation to begin soon after sunset, the wind speed t 2 m seldom dropped below 0·5 m/sec. When filter-paper absorption showed a steady ccretion of moisture on the leaves with little or no weight increase of the balance pan, wo conditions must have obtained; (i) turbulent transfer of water vapour from the air

must have been extremely small and (ii) the entire flux of vapour from the soil must have condensed on the leaves. In other words, the amount of water vapour carried to the surface by turbulence on very calm nights was insignificant compared with the amount of vapour reaching the grass by diffusion from the soil. With increasing wind strength, the turbulent flux increased, reaching a maximum observed value of 3·5 mg cm^{-2} hr^{-1} on a night when the wind speed at 2 m was 2-3 m/sec.

It is proposed that the two sources of condensation on a short grass surface should be distinguished by the names ' distillation ' (for the upflux from the soil) and ' dewfall ' (for the downflux from the atmosphere).*

5. Distillation

The regular increase in the weight of moisture absorbed by filter papers (Figs. 6, 7, 8) gave confidence in the technique as a relative measurement of condensation. It was impossible to estimate how much moisture was left behind on the leaves when a sample was taken, but on the evidence of one night's observations (Fig. 8) when the dew balance recorded twice the condensation absorbed by filter papers it appears that this may have been as much as 50 per cent. On nights when there was considerable distillation but little dewfall, absorbed moisture measurements can be compared with the rate of distillation D calculated from Eq. (1), assuming $K_v = 0{\cdot}24$ cm^2/sec. Since the air was always within one or two per cent of saturation it was convenient to work in terms of temperature differences, writing

$$D = K_v \{T(0) - T(1)\} \left(\frac{\partial x}{\partial T}\right)_s \quad . \quad . \quad . \quad . \quad (3)$$

where $(\partial x/\partial T)_s$ is the rate of change with temperature of the saturated absolute humidity calculated at the mean of the surface temperature, $T(0)$, and the air temperature at 1 cm, $T(1)$.

TABLE 3. Distillation data

Date	Time of moisture sampling GMT	Mean rate of measured-moisture increase (mg cm^{-2} hr^{-1})	Time of temp. observation GMT	$T(0)$ (°C)	$T(1)$ (°C)	Distillation (Eq. 3) (mg cm^{-2} hr^{-1})
3 Oct. 1953	1745	—				
			1800	12·0	8·9	1·7
	1845	1·2				
			1900	10·9	7·6	1·7
	2045	1·6				
			2000	9·2	7·7	0·7
	2245	1·2				
5 Oct. 1953	1845	1·0				
			1900	8·5	4·7	1·7
	1945	1·1				
			2000	8·1	5·0	1·4
	2045	0·8				
			2100	7·4	3·0	1·8
	2145	2·0				
			2200	6·9	2·1	1·9
9 Oct. 1953	1815	—				
			1900	6·9	3·2	1·5
	1915	1·0				
			1000	5·7	2·1	1·3
	2015	0·5				
			2100	5·5	2·2	1·2
	2115	0·9				

* It must be emphasized that the observations described here were not complicated by the occurrence of guttation, the exudation of a water droplet at the tip of a blade of grass, which is readily distinguishable from condensation (Long 1957).

Simultaneous observations of moisture and temperature were impossible, but in Table 3, mean rates of moisture increase for one or two-hour sampling on three typical nights are compared with distillation rates estimated from Eq. (3), and 10 min. mean values of temperature. In view of the uncertain collection efficiency of the filter papers and of the possibility that neither the temperature nor the moisture values were perfectly representative of the surface as a whole, agreement is reasonable. For all three nights, observed and calculated rates of distillation were of the order of 1-2 mg cm^{-2} hr^{-1}, and the mean ratio for 10 observations of observed/calculated distillation was 0·76. Further error may have arisen from the assumption of molecular transfer, of linear temperature gradient, and of saturation at the soil surface. Nevertheless, the figures in Table 3 lend further support to the thesis that water vapour diffuses upward from the soil on calm nights and that the magnitude of the flux can be estimated, to within a factor of two at least, from simply observed gradients of temperature and humidity.

6. Dewfall

In the restricted sense of this paper, dewfall is the eddy flux of water vapour from the atmosphere to the surface and may therefore be subjected to the same analysis which has previously been applied to evaporation. Observations by Rider (1954) suggest first, that it is justifiable to assume identity of the eddy coefficients for water vapour and momentum in stable conditions, and second, that the dependence of these coefficients on stability, height, and wind speed may be expressed by an extension of the equation for neutral conditions as proposed by Rossby and Montgomery (1935) to give

$$K = k^2 z^2 \frac{\partial u}{\partial z} \Big/ (1 + \sigma Ri) \quad . \quad . \quad . \quad . \quad (4)$$

where σ is a number assumed independent of height and stability, k is von Kármán's constant = 0·4 and $\partial u/\partial z$ is the wind gradient at height z. The Richardson number or stability parameter is defined with sufficient precision for the present observations as

$$Ri\,(z) = \frac{3{\cdot}5\,(\partial T/\partial z)}{(\partial u/\partial z)^2} \quad . \quad . \quad . \quad . \quad . \quad (5)$$

From measurements of shearing stress, wind and temperature gradients, Rider obtained values of σ from 8·0 to 9·1 while Pasquill (1949b) working from evaporation measurements and the assumption that transfer coefficients of momentum and water vapour are equal, found $\sigma \simeq 12$.

Eqs. (1) and (4) lead to an expression for the downward flux of water vapour

$$F = \frac{k^2 z^2 (\partial u/\partial z)(\partial x/\partial z)}{1 + \sigma Ri} \quad . \quad . \quad . \quad . \quad (6)$$

All quantitites in Eq. (6) except σ could be measured directly and a determination of σ with a check on its independence of stability should therefore have been possible from the present observations. But when Eq. (6) is used in this way errors in measured flux and in gradients produce very much larger errors in σ. If fractional errors are written in the form $\frac{\Delta\sigma}{\sigma}$ then

$$\frac{\Delta\sigma}{\sigma} = \left(1 + \frac{1}{\sigma Ri}\right)\frac{\Delta\theta}{\theta} - \frac{\Delta Ri}{Ri}$$

where

$$\theta \propto \frac{1}{F}\frac{\partial u}{\partial z}\frac{\partial x}{\partial z}$$

For errors of ± 10 per cent in each of $\partial T/\partial z$, $\partial u/\partial z$ and $\partial x/\partial z$ (unavoidable in graphica determinations of gradients, see Rider 1954) and of ± 10 per cent in F, $\frac{\Delta\theta}{\theta} = 0{\cdot}17$ an $-\frac{\Delta Ri}{Ri} = 0{\cdot}1$. Hence with $\sigma = 10$, $\frac{\Delta\sigma}{\sigma} = 2{\cdot}0$ with $Ri = 0{\cdot}01$, and $\frac{\Delta\sigma}{\sigma} = 0{\cdot}44$ wit $Ri = 0{\cdot}1$. Within the normal range of nocturnal stability, individual determinations of are therefore liable to considerable error and its constancy could be established only many observations were available of fluxes and gradients at different stabilities. Thes were not provided by the present observations, partly because dewfall nights were s rare and partly because of the unexpectedly erratic nature of the humidity profile Smooth profiles of temperature were obtained, however, and when the air was almos saturated, with wet- and dry-bulb temperatures virtually equal at all heights, reliabl humidity gradients could be derived. From several hundred ten-minute runs, 11 wer eventually found in which dewfall was at least 0·4 mg cm^{-2} hr^{-1}, relative humidit between 99 and 100 per cent, and wind speed at the lowest observation point (25 cm was greater than the stopping speed of the anemometers. Eq. 6 was then used to comput F assuming $\sigma = 10$, the mean of Pasquill's and Rider's determinations (Table 4). Fro Fig. 9 in which observed and computed dewfall are compared, it appears that dewfa over a short grass surface in saturated air can be estimated with an error of approximatel ± 30 per cent if wind and temperature gradients are known. The scatter of points abou the line of unit slope is a measure of the apparent deviation of σ from its assumed value c 10. Values of σ computed from Eq. (6) and observed values of F ranged from 0 to 55 wit a mean of 12.

TABLE 4. CALCULATION OF DEWFALL $F = k^2 z^2 \frac{\partial u}{\partial z}\frac{\partial x}{\partial z}\Big/(1 + 10\,Ri)$

Date	Time GMT	u (50) (cm/sec)	T (50) (°C)	$\partial u/\partial z$ (sec^{-1})	$\partial T/\partial z$ (°C/m)	$\partial x/\partial z$ (g/m^4)	$Ri \times 10^3$	Dewfall (mg cm^{-2} hr^{-1}) Computed from Eq. (6)	Observe
1 Sept. 1953	0300	121	11·20	0·56	0·62	0·43	69	2·1	2·7
	0400	288	12·38	0·84	0·22	0·16	11	1·7	1·2
29 Sept. 1953	0330	71	4·52	0·42	1·0	0·47	200	0·95	1·1
	0400	100	5·09	0·50	0·67	0·33	94	1·2	1·2
	0530	89	5·02	0·49	0·50	0·24	72	0·98	1·5
2 Oct. 1953	0100	95	9·67	0·60	0·90	0·57	87	2·6	2·1
	0200	108	9·75	0·57	0·55	0·35	59	1·8	2·9
	0230	99	9·27	0·61	0·73	0·46	69	2·4	2·7
27 Apr. 1954	0315	245	2·33	0·79	0·25	0·10	14	1·0	0·97
	0345	218	2·21	0·65	0·21	0·086	17	0·69	0·75
	0415	207	2·14	0·66	0·19	0·076	15	0·63	0·61

Holzman (1943) proposed another form of equation for vapour transfer in non-neutra conditions in which the factor $1/(1 + \sigma Ri)$ in the Rossby-Montgomery formulatio was replaced by $(1 - \sigma' Ri)$ to facilitate integration of the wind-profile equation. Previou work summarized by Deacon (1956) gives values of σ' ranging from 7·1 to 12 fo $-0{\cdot}33 < Ri < +0{\cdot}08$. An attempt to fit present observations to the Holzman equatio gave an even greater variation in σ' than was found in σ and the existence of appreciabl water-vapour flux with Ri as high as 0·2 suggests that the Holzman form is unsuitabl for extrapolation to conditions of extreme stability.

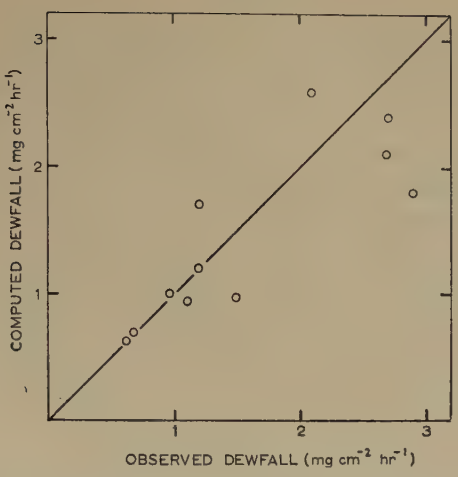

Figure 9. Dewfall observed on balance and computed from Eq. (6) with line of unit slope.

Neumann (1956) has adopted Eq. (6) with $\sigma = 10$ in order to estimate dewfall from observations of temperature, humidity and wind at screen level, and from the reading of a grass minimum thermometer. Predicted dewfall is of the right order but further work along the same lines should aim at modifying the assumption that temperature and humidity gradients, and hence dewfall, are constant throughout the night. Figs. 7 and 8 illustrate characteristic dewfall nights in which a short period of relatively constant rate of dewfall in the early morning was preceded by a long period of slowly increasing rate.

This slow increase of dewfall rate has an important bearing on the dewfall total for a night. Between the transition and the condensation regimes, when the atmospheric water-vapour flux is zero, the vapour-pressure gradient near the surface must also be zero. If the air in contact with the surface is saturated, its vapour pressure is uniquely determined by the surface temperature and the subsequent development of the vapour-pressure inversion required for dewfall depends on the rate of surface cooling, which is greatest at sunset. Observed dewfall rates increased most rapidly when the humidity of the air in the late afternoon was sufficiently high to permit early condensation. When condensation was delayed long after sunset on account of low humidity (e.g., on 26 April 1954), total dewfall was always very small. A comparison of dewfall and weather on 5 nights is given in Table 5.

TABLE 5. WEATHER DATA FOR DEWFALL NIGHTS

ΔT = temperature — dew point at sunset; Δt = time after sunset when dew first recorded

Date	26-27 April 1954	31 Aug.-1 Sept. 1953	28-29 Sept. 1953	1-2 Oct. 1953	3-4 Oct. 1953
Approximate 2 m wind (m/sec)	3-4	3-4	1-1·5	1-1·5	1-1·5
ΔT (°C)	9·0	3·2	4·3	4·5	4·2
Δt (hr)	6·8	3·2	1·8	1·8	1·0
Max. dewfall (mg cm^{-2} hr^{-1})	1·2	3·5	2·0	3·0	0·6
Total dewfall	3	13	11	12	3·5*

* dewfall interrupted 7 hr after sunset by development of complete cloud cover.

Present observations are supported by those of Frankenberger (1955) who found that on clear nights maximum dewfall on a meadow (2·4 mg cm^{-2} hr^{-1}) was obtained when the wind speed at 10 m was 3-4 m/sec. With wind speed 0-1 m/sec dewfall was only 0·3 mg cm^{-2} hr^{-1} but with wind in excess of 5 m/sec, evaporation was observed. The rarity of dewfall nights at the present site can be attributed to the difficulty of obtaining precisely the right combination of humidity, wind speed and cooling rate. A clear sky was essential for rapid cooling; optimum conditions at screen height were a relative humidity of at least 75 per cent at sunset and a wind of 1-3 m/sec maintained throughout the night. Such a wind is uncommon on clear nights at all times of year since the development of a temperature inversion checks the exchange of momentum with upper levels. Nocturnal relative humidities, however, show a marked seasonal variation. Maximum humidities occur in the autumn when land is cooling faster than surrounding sea; and since long anticyclonic spells with cloudless nights are more frequent between mid-September and mid-October than at any other time of year, this is the season most favourable for dewfall. Cloudless weather in spring is often unsuitable for dewfall since stationary anticyclones normally bring dry air from the continent.

From the observed frequency of dewfall nights, it seems unlikely that there should be more than 20 to 25 nights in a normal year when dewfall reaches 10 mg/cm^2. On this basis, the annual dewfall on a similar short-grass surfaces in the British Isles may be estimated as 0·1-0·2 in. This is much lower than previous estimates based on the amount of condensation on artificial collecting surfaces, e.g., 1-1·5 in. (Dines 1879); 0·9 in. (Skinner 1912). Annual distillation must be considerably greater than dewfall but since distillation is simply the transfer of moisture from one part of the surface to another, it has no hydrological significance.

7. The heat-balance equation

In preceding sections, dewfall and distillation have been regarded as fluxes of water vapour. They may also be regarded as fluxes of latent heat, distillation representing heat transfer from soil to leaves, and dewfall, heat transfer from atmosphere to leaves. Further light can be shed on the relative importance of these two processes by examining the heat exchanges which take place within the cover. The heat-balance equation for the centimetre-deep layer of grass and air immediately above the soil surface can be written

$$R + G + C + \lambda F + M = Q\frac{\partial T}{\partial t} \quad . \quad . \quad . \quad (7$$

where

R = net *incoming* radiation flux (i.e., a negative term)
G = flux of total heat from soil to surface
C = flux of sensible heat from air to surface
F = flux of water vapour from air to surface
λ = latent heat of vaporization
Q = heat capacity of the air-grass layer per cm^2
T = mean temperature of layer
M = heat released by vegetable metabolism per cm^2

Eq. (1) involves no explicit term for distillation since the measured flux G is the sum of sensible and latent heat transfer from the soil.

The terms of the equation will now be discussed individually.

(*a*) *Radiative flux R*

From the measured variation of temperature and humidity with height up to 2 m combined with the upper-air data from a suitable radiosonde ascent, it is possible to construct a radiation diagram (Robinson 1950) from which the net radiative loss at night when the sky is cloudless may be calculated. Groen (1947) has discussed the decrease in net outgoing radiation during the night due to the fall of temperature and the decrease in vapour pressure near the surface. For present purposes this decrease could be ignored. The change in downward flux caused by the change in water path-length was very small and temperature dependence was unimportant because the decrease in the upward flux related to surface cooling was largely offset by the decrease in downward flux related to air cooling. The computed radiation at 0200 GMT could therefore be adopted as a constant dusk-to-dawn value provided the sky remained clear, the error involved being much less than error inherent in the method itself.

From Robinson's (1947) paper an uncertainty of about $\pm$ 5 per cent in downward radiation may be inferred owing to additional radiation from atmospheric pollution or from very high tenuous cloud. Since net flux at the surface is of the order of a quarter of downward flux, the error in calculated net flux could amount to $\pm$ 20 per cent.

On each occasion, the ascent from Crawley at 0200 GMT was used for the upper-air data. Crawley lies 50 mi SW of Harlington, Middlesex, where the observations were made, and in the absence of frontal surfaces no significant difference would be expected in the conditions above, say, 900 mb at the two stations. The form of the radiation diagram *below* 900 mb was indicated with adequate precision by the surface observations at Harlington owing to the small variation of the fourth power of absolute temperature between the 900 mb level and a height of 2 m above the surface.

(*b*) *Soil-heat flux G*

The main component of soil-heat flux was measured with a flux plate of the type described by Deacon (1951) consisting of a sheet of Paxolin 9·5 × 2 × 0·16 cm with 60 copper-constantan thermocouples on each face. The plate was calibrated at Kew Observatory in a large box containing moist soil and provided with a heating system such that a known flux of heat could be passed through the plate. The sensitivity was found to be 62 μV per cal cm^{-2} hr^{-1} and experience at Kew suggested that this figure should not differ by more than about $\pm$ 10 per cent for completely dry and completely saturated soil. The plate was buried at a depth of 4 cm beneath turf at the experimental site and was allowed to settle down for 10 days before taking readings.

The flux from the layer between 4 cm and the surface was obtained from temperature measurements at these two levels. A graph of the mean temperature against time was used to determine the rate of temperature change and the flux was then estimated as $4\rho_s c_s \frac{\partial T}{\partial t}$ assuming constancy in the 4 cm layer of the apparent density ρ_s and the apparent specific heat c_s which were derived from the moisture content and apparent density of samples.

The specific heat of the soil was not determined directly, but from data given by Keen (1931) the appropriate value for a clay, 0·22, was selected. The method used to determine apparent heat capacity does not justify the use of two significant figures and a value of 0·8 cal/cm^3 was found when the soil was near field capacity.

(c) *Sensible heat flux C*

The convective flux of heat was the least accurately determined quantity in the balance. In the first place it was necessary to assume that the horizontal advection of heat was negligible. In fact, calm nights are characterized by temperature fluctuations, wholly or partly due to advection, whose periods are of the same order as the observation period (Rider and Robinson 1951). In the second place, the formula used to compute C has only slender empirical justification.

Observations by Pasquill (1949b) suggest equality of the transfer coefficients for water vapour and heat over grass in stable conditions, while Rider's (1954) results support equality of the coefficients for momentum and vapour transfer. Until further evidence accumulates it may be assumed that all three coefficients are equal at night in the presence of a temperature inversion and that their dependence on stability is given by Eq. (4) with $\sigma = 10$. The heat flux is then calculated as

$$C = \rho c_p K \frac{\partial T}{\partial z} \quad . \quad . \quad . \quad . \quad (8)$$

where ρ and c_p are respectively the density of air and its specific heat at constant pressure.

The gradients $\partial T/\partial z$ and $\partial u/\partial z$ are difficult to determine accurately and for this reason alone C could be in error by $\pm$ 20 per cent. On many occasions the wind gradient could not be measured owing to stopping of the anemometers in lights winds but since in these cases deliberate overestimates of $\partial u/\partial z$ gave C less than 0·4 cal cm^{-2} hr^{-1}, convective heat transfer could be ignored. Apart from these practical difficulties, the application of Eq. (3) to the determination of heat flux at times when the temperature gradient may have been controlled by radiative transfer is a dubious procedure. Fortunately these were times at which C played a very small part in the heat balance.

(d) *Latent heat flux λF*

Latent heat transfer from atmosphere to surface was estimated directly from the weight increase recorded by the dew-balance and the error in this term can be neglected compared with those already discussed.

(e) *Ratio of sensible and latent heat flux*

It will be convenient to derive an expression for the ratio of the fluxes of latent and sensible heat transfer in a saturated atmosphere. Assuming identity of the transfer coefficients, this is

$$\frac{\lambda F}{C} = \lambda \frac{\partial x}{\partial z} \Big/ \left(\rho c_p \frac{\partial T}{\partial z} \right)$$

where x is absolute humidity and T air temperature at height z.

Since

$$\frac{\partial x}{\partial z} = \rho \epsilon \frac{\partial T}{\partial z} \frac{\partial e}{\partial T} \Big/ p$$

where e and p are saturated water vapour and total atmospheric pressures in the same units, and ϵ is the ratio of the densities of water vapour and dry air at the same pressure and temperature, it follows that

$$\frac{\lambda F}{C} = \frac{\partial e}{\partial T} \Big/ Ap \quad . \quad . \quad . \quad . \quad . \quad (9)$$

where A is the ' constant ' of the psychrometer equation, $c_p/\lambda\epsilon$.

The ratio increases from 0·7 at 0°C to 2·2 at 20°C. For an unsaturated atmosphere with relative humidity decreasing with height, a common nocturnal situation, it can readily be shown that the ratio $\lambda F/C$ is less than the value for saturated air. Hence if dewfall is known, Eq. (9) can be used to find the minimum sensible heat flux associated with the vapour transfer, whatever the relative humidity of the air.

(*f*) *Heat of metabolism M*

The contribution of metabolism to the nocturnal heat balance is usually ignored or dismissed as negligible but no quantitative estimates have been found in the literature. Metabolic heat is produced in the roots and leaves of plants by the oxidation of sugars

$$C_6 H_{12} O_6 + 6 O_2 \rightarrow 6 CO_2 + 6 H_2O + 673 \text{ Kg cal}$$

the production of 1 g CO_2 being accompanied by the release of 2·4 Kg cal. The available evidence (James 1953) suggests that practically all this heat is available as a contribution to the heat budget of the surroundings.

Measurements by various workers in the laboratory (Richards 1938, Woodford and Gregory 1948) gave respiration rates of 2·1 mg CO_2/g dry weight for leaves and 5 mg CO_2/g dry weight for roots, both at 10°C. Turves cut at Harlington in the spring and a few after cutting of the grass gave 4 to 5 $\times 10^{-2}$ g dry weight of vegetable matter per cm^2 of soil divided in roughly equal parts between leaves and roots. A reasonable estimate of CO_2 production is therefore 0·2 mg cm^{-2} hr^{-1}, which would be associated with the release of 0·5 cal cm^{-2} hr^{-1}. In the present study this quantity is of the same order as the error involved in the measurement of other terms in the heat balance and it may therefore be ignored. For long grass and other denser types of cover, or when net radiation is small owing to the presence of cloud, metabolism may make a significant contribution to the nocturnal heat balance.

(*g*) *Change of heat content*

The value of Q estimated from fresh weight of leaves was 0·2 cal/cm^3 and $\partial T/\partial t$ was of the order of 1°C/hr on most nights. Changes in the heat content of the surface layer were therefore insignificant.

8. Analysis of records

(*a*) *Heat balance on very calm nights*

Table 6 contains heat-balance data for 3 nights (i, ii, iii) when there was seldom sufficient wind to rotate the anemometers and when estimated sensible heat transfer was negligible. No dewfall was recorded, but slight evaporation persisted throughout the night of 7-8 May 1954, at a rate between 0·9 and 0·3 mg cm^{-2} hr^{-1}. An instrumental defect might have been suspected but an upward flux of water vapour was confirmed by vapour-pressure observations. For example, at 2200 GMT, vapour pressures at 12·5 and 100 cm were 7·71 and 7·09 mb respectively. On nights (i) and (ii) the values of total soil-heat flux (G) and the radiation loss ($-R$) are remarkably close but in view of the possible error in both quantities the agreement must be regarded as fortuitous. On night (iii) there is a gradually decreasing discrepancy between radiation loss and the sum of the remaining terms in the heat balance.

TABLE 6. HEAT-BALANCE COMPONENTS

G_1 = heat flux from ground below 4 cm as recorded by flux plate

G_2 = heat flux from ground above 4 cm $= 4\,\rho_s\,c_s\,\frac{\partial T}{\partial t} = 3{\cdot}2\,\frac{\partial T}{\partial t}$

For other symbols see § 7. All heat fluxes are in cal cm^{-2} hr^{-1}

(i) 6-7 May 1954
$R = -8{\cdot}3$; $\lambda F = C = 0$

Time (GMT)	$T(0)$ (°C)	$T(-4)$ (°C)	G_1	G_2	G
1900	8·7	10·3	2·0	3·8	5·8
1930	7·9	10·0	3·6	3·8	7·4
2000	7·1	9·5	4·9	3·8	8·7
2030	6·6	9·1	5·6	3·6	8·2
2100	6·4	8·8	6·0	1·9	7·9
2130	6·0	8·5	6·2	1·9	8·1
2200	5·8	8·2	6·4	1·9	8·3
2230	5·4	7·9	6·4	1·9	8·3
2300	5·2	7·6	6·6	1·6	8·2
0000	4·7	7·2	6·9	1·6	8·5
0100	4·3	6·8	6·7	1·6	8·3
0200	3·8	6·1	7·2	1·3	8·5
0300	3·3	6·0	7·5	1·0	8·5

(ii) 11-12 May 1954
$R = 6{\cdot}5$

Time GMT	$T(0)$ (°C)	$T(-4)$ (°C)	G_1	G_2	C	λF	$G+C+\lambda$
1900	17·4	18·7	0·7	4·5	0·8	− 2·9	3·1
2000	15·6	17·6	2·1	3·8	0	0	5·9
2030	15·1	17·2	2·5	2·9	0	0	5·4
2100	14·6	16·9	2·9	2·9	0	0	5·8
2130	13·8	16·4	3·6	2·9	0	0	6·5
2200	13·5	16·1	4·0	1·9	0	0	5·9
2300	12·8	15·4	4·3	2·2	0	0	6·5
0000	12·1	14·7	4·3	1·9	0	0	6·2
0100	11·5	14·2	4·5	1·9	0	0	6·4

(iii) 7-8 May 1954
$R = -7{\cdot}2$; $C = 0$

Time GMT	$T(0)$ (°C)	$T(-4)$ (°C)	G_1	G_2	λF	$G + \lambda F$
1930	9·4	11·7	3·6	3·8	− 0·5	6·9
2000	8·7	11·2	4·7	3·5	− 0·5	7·7
2030	8·1	10·7	5·7	3·2	− 0·5	8·4
2100	7·1	10·3	6·0	2·9	− 0·5	8·4
2130	7·4	10·0	5·9	2·6	− 0·5	8·0
2200	6·9	9·5	6·5	2·6	− 0·5	8·6
0100	4·9	7·7	7·5	1·9	− 0·2	9·2
0200	4·2	7·3	7·7	1·4	− 0·2	8·9
0300	4·0	6·9	7·9	1·3	− 0·2	9·0
0400	3·6	6·5	7·9	1·0	− 0·2	8·7
0500	3·6	6·1	7·4	0·6	− 0·4	7·6

(iv) 10-11 May 1954
$R = -6{\cdot}8$

Time (GMT)	$T(0)$ (°C)	$T(-4)$ (°C)	G_1	G_2	C	λF	$G+C+\lambda F$
1800	19·4	19·6	− 1·3	3·8	1·6	− 6·5	− 2·4
1900	17·9	18·3	0·2	3·8	3·5	− 5·8	1·7
2000	16·6	17·3	1·6	3·5	1·5	− 3·1	3·5
2100	15·1	16·1	2·8	3·2	1·5	− 2·0	5·5
2200	—	—	3·5	2·9*	1·4	− 1·3	6·5
2300	13·1	14·3	3·9	2·2	0·4	− 0·4	6·1
0330	11·8	12·9	4·4	1·6	0·2	0·0	6·2
0400	10·5	11·6	4·8	1·0	0·4	0·0	6·2
0430	10·5	11·5	4·5	0·6	1·0	0·0	6·1
0500	10·8	11·6	4·2	0·3	0·5	0·0	5·0

(v) 1-2 Oct. 1953
$R = -6{\cdot}5$; G estimated as $-(R + C + \lambda F)$

Time (GMT)	C	λF	G
1730	3·1	− 2·9	6·3
1800	3·1	− 1·4	4·8
1830	2·9	− 1·4	5·0
1900	1·6	− 0·5	5·4
1930	1·4	0·1	5·0
2000	0·7	0·2	5·6
2030	0·9	0·3	5·3
2100	3·6	0·3	2·6
2130	2·1	0·6	3·8
2200	2·5	0·6	3·4
0000	1·9	0·8	3·8
0100	1·2	1·2	4·1
0200	0·8	1·7	4·0
0230	1·1	1·6	3·8

* estimated

One feature common to all 3 nights was the deposition of moisture on the leaves, attributed to upward distillation since the balance showed no weight increase. At 0300 GMT on 7 May 1954 filter-paper absorption of this moisture gave a mean distillation rate for the previous 8 hr of 1·5 mg cm^{-2} hr^{-1} and the nature of the technique makes this a minimum estimate. Had this moisture been deposited by downward transfer from the

atmosphere, the associated latent heat flux would have been 0·7 cal cm^{-2} hr^{-1} and since the surface temperature at 0100 GMT was approximately 4°C the associated sensible heat flux would have been at least 0·96 cal cm^{-2} hr^{-1} (Eq. 9). Hence if condensation were to be attributed to dewfall rather than to distillation, an additional 1·8 cal cm^{-2} hr^{-1} would be introduced to the system and agreement between radiative loss and the positive terms in the heat balance would become very much poorer. At the same time, it could be argued that since the present agreement is fortuitously close, a discrepancy of the correct magnitude and sign might be concealed by error in R and G. But on 11-12 May 1954 radiative loss and soil heat flux were again nearly equal and again the heat balance would be upset by the addition of terms representing atmospheric heat transfer. On 6-7 May 1954 a poor balance would be made still worse by the addition of a positive term. Moreover, estimates of sensible heat transfer, albeit with an empirical equation and with uncertain wind profiles, gave estimates of $C < 0{\cdot}4$ cal cm^{-2} hr^{-1}.

Heat-balance estimates therefore provide further evidence that the moisture condensing on short grass on very calm nights represents a transfer of latent heat to the grass, not from the atmosphere, but from the soil surface.

It may be noted that while the assumption of molecular exchange within the cover leads to a figure of 1·8 cal cm^{-2} hr^{-1} for the total heat transfer from soil to leaves, the measured heat loss from the soil was of the order of 8 cal cm^{-2} hr^{-1}. This discrepancy could arise from underestimation of the exchange coefficient. But it is possible that approximately 6 cal cm^{-2} hr^{-1} were lost by direct radiation to the atmosphere from the soil surface, in other words, that the short-grass cover of a playing field is a relatively inefficient distillation system.

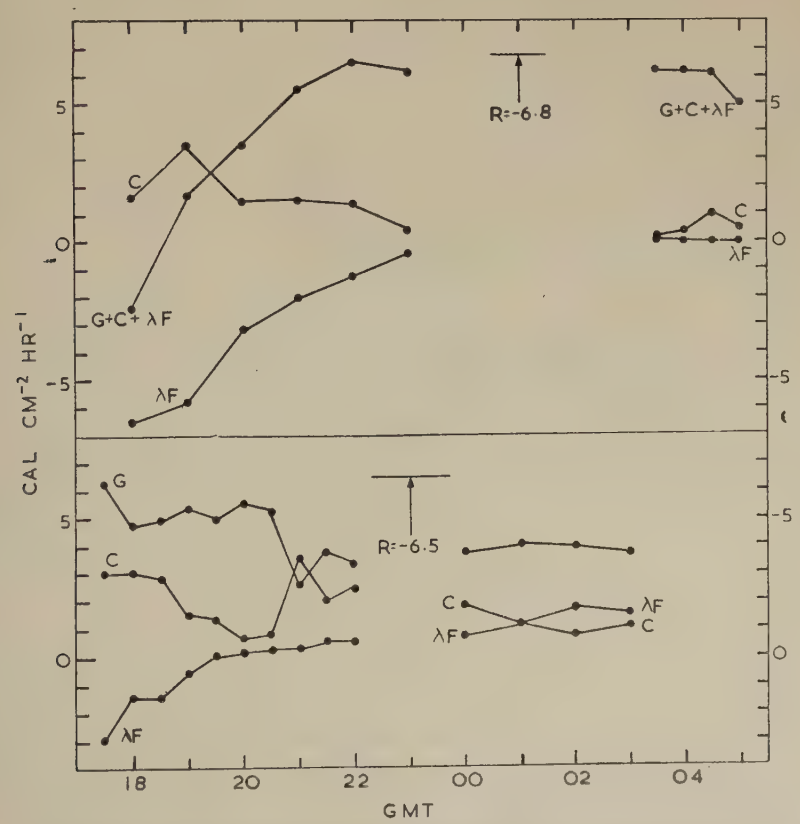

Figure 10. Heat-balance components of a short-grass surface: upper; 10-11 May 1954: lower; 1-2 Oct. 1953.

(b) Heat balance with sensible and latent heat transfer from the atmosphere

On the night of 10-11 May 1954, the mean wind speed at 2 m varied between 1 and 2 m/sec and the estimated sensible heat transfer from the atmosphere was appreciable. Dew-balance records showed that evaporation took place till 0200 GMT. The complete heat balance (Fig. 10 and Table 6 (iv)) again shows good agreement between radiative loss and the sum of the positive terms. The decrease in sensible heat flux was balanced by decreasing evaporation so that the heat flux from the ground varied little through the night.

Unfortunately, dewfall was extremely slight on the nights in the spring of 1954 when soil heat flux was directly measured, but since the heat balances for these nights were encouraging, a similar analysis was performed for nights in the autumn of 1953 when dewfall was considerable but when it was necessary to estimate G as $-(R + C + \lambda F)$. A striking feature of the example given in Fig. 10 and Table 6 (v) is the decrease in sensible heat transfer till 2000 GMT, followed by a marked increase at 2100 GMT as the result of a sharp decrease in stability. Associated with the change in atmospheric flux, the estimated soil flux changed abruptly but showed little change after 2100 GMT. The condensation rate increased throughout the night and the latent heat transfer λF finally exceeded the sensible heat transfer C. At 0200 GMT the surface temperature was about 9°C giving a theoretical ratio of 1·2 for $\lambda F/C$ which is roughly the observed magnitude. It would appear that when wind speed is constant, the *total* atmospheric heat flux is also approximately constant so that as dewfall increases, convection must decrease.

9. Conclusion

The observations have made it possible to distinguish two types of dew formation on one particular type of cover in southern England. Implications for other crops and other climates may very briefly be considered.

Measurements of the soil-heat flux beneath potatoes and beneath wheat (Monteith 1957) showed that with wet soil and low wind, conduction through the soil and radiative loss were nearly equal at night, implying that atmospheric fluxes were negligible and that condensation should be attributed to distillation from the soil. Crops taller and denser than short grass can shield the soil from direct radiative loss and if the entire heat loss from the soil were by sensible and latent heat transfer to the cover, distillation rates could approach 7 to 8 mg cm^{-2} hr^{-1} in English summer or autumn weather. In short grass on calm nights moisture transfer occurs in a quasi-laminar boundary layer of height comparable with that of the grass itself. Within crops however, temperature and humidity gradients are much smaller (see e.g., Hirst, Long and Penman 1954) and, for a given heat flux, transfer coefficients within the cover may therefore be much larger than in short grass. The stalks and leaves of crops may project well above the region of laminar flow even on calm nights and may therefore collect more dewfall from the atmosphere than would short grass in similar weather (Long 1957).

In tropical and sub-tropical countries, with the possibility of higher nocturnal temperatures and a greater radiative loss at night than in Britain, at least one present conclusion should still be valid: dewfall on short grass should be negligible when surface-wind speeds are of the order of 50 cm/sec or less. But with temperature minima of 20-25°C the temperature-dependent ratio of latent to sensible heat transfer in the atmosphere will be two to three times more than on a typical summer or autumn night in Britain and dewfall, when it does occur, will be correspondingly greater.

Acknowledgments

I should like to express my gratitude to Dr. H. L. Penman for suggesting that I should undertake this work and for much helpful discussion subsequently; to Dr. B. J. Mason who supervised the investigation at Imperial College; and to the Department of Scientific and Industrial Research and the Agricultural Research Council for research grants. Part of the material is taken from a Ph.D. thesis submitted to the University of London.

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The thermodynamics of saturated air : a contribution to the classical theory

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SUMMARY

The temperature attained in the reversible and pseudo-adiabatic expansion of air saturated with respect to liquid water is taken as a reference; the temperatures attained in the following expansions from the same initial state are then compared with it : (a) the ice-saturation pseudo-adiabatic process (b) the water-saturation adiabatic process (c) the ice-saturation adiabatic process.

The results of these comparisons are summarized in Figs. 1 to 5 of this paper.

Similarly, the cloud virtual temperature, the virtual temperature adjusted for the weight of suspended particles, attained in water-saturation adiabatic expansion, is compared with that attained in water-saturation pseudo-adiabatic expansion; Fig. 6.

The implications of these calculations for the properties and behaviour of cumulus are commented upon.

1. INTRODUCTION

Both observational and theoretical studies of cumulus have emphasized the importance role that cloud temperature plays in determining the strength of cloud updraughts. Apart from any intrinsic importance, updraughts together with cloud thickness, control the lifetime of particles within cloud and thus help to determine whether precipitation forms or not.

Some success has been achieved in deducing cloud temperature (and droplet and crystal content too) from a simple thermodynamic theory probably first proposed by Guldberg and Mohn (1876). In this classical approach, the cloud properties at any level are deduced on the assumption that cloud air elements ascend to that level from cloud base, the elements undergoing only adiabatic and reversible changes of state.

These hypotheses, however, do not define uniquely the properties of expanding saturated air. In addition, the nature of the cloud particles must be specified, since at sub-zero temperatures vapour may condense on to either crystals or droplets. Furthermore an adiabatic process must be distinguished from a pseudo-adiabatic process (von Bezold 1888); whilst in the former process the condensation products are retained in the ascending element of air, in the latter process they are supposed precipitated from it. The implications of these alternatives – ice or supercooled water, adiabatic or pseudo-adiabatic – are not always clearly appreciated.

Measurements from aircraft of temperature in cloud unfortunately do not possess the desired reliability because of the very considerable instrumental difficulties, and no measurement of cloud ice-crystal content has been published. The liquid-water content of cloud, however, has been measured extensively and comparisons have been made between observed values and those deduced on the hypothesis of adiabatic and reversible ascent of saturated air from cloud base, the condensate supposed liquid water throughout. These observations (Zaitsev 1950, Weickmann and aufm Kampe 1953, Murgatroyd 1955, Warner 1955) reveal that, whilst on the majority of occasions the theoretical value is a large over-estimate, on some occasions and especially in cumuliform cloud the theoretical value is approached. (The importance of peak values will be appreciated if it is recognized that the regions of high water content favour those droplet aggregation processes which can lead to the formation of precipitation).

Because, then, the classical estimate of cloud properties is sometimes adequate and also because any other theoretical estimate must essentially be a refinement of the

classical treatment, it is important to understand clearly what the classical theory implies. This has been attempted in the following pages; in particular the temperatures attained in reversible adiabatic and pseudo-adiabatic expansions, in which air is saturated either with respect to liquid water or with respect to ice, have been compared.

Such comparisons have been made previously by Neuhoff (1900) and Fjeldstad (1926). Neither author, however, has considered the possibility of supercooling; that is, both assumed that at temperatures below 0°C any condensate is of ice. Furthermore, both have expressed their results as differences in pressure at the same temperature; thus "air initially saturated at a temperature of 20°C and a pressure of 760 mm reaches a temperature of −18°C at a pressure of 316 mm in the pseudo-adiabatic system, and 304 mm in the adiabatic system."

Comprehensive and more convenient comparisons (of temperature at the same pressure) are described below.

2. List of symbols

C_p = Specific heat at constant pressure for dry air 0·240 IT cal gm^{-1} °K^{-1}

C_p' = Specific heat at constant pressure for water vapour 0·441 IT cal gm^{-1} °K^{-1}

C_i = specific heat of ice 0·5 IT cal gm^{-1} °K^{-1}

C = specific heat of water 1·0 IT cal gm^{-1} °K^{-1}

L_f, L_s, L_v = latent heat of fusion, sublimation, vaporization

r_w = saturated mixing ratio with respect to water

r_i = saturated mixing ratio with respect to ice

r_v = mixing ratio, vapour to dry air

r_{cw} = mixing ratio, liquid water condensate to dry air

r_{ci} = mixing ratio, ice condensate to dry air

r = total mixing ratio, all phases of water to dry air

T = temperature on absolute scale, T_v virtual temperature

ρ = (moist) air density

p = total atmospheric pressure

p_a = partial pressure of dry air

e_w = saturation vapour pressure over water

e_i = saturation vapour pressure over ice

R = specific gas constant for dry air $6{\cdot}8557 \times 10^{-2}$ IT cal gm^{-1} °K^{-1}

ϵ = ratio between the densities of water vapour and dry air (0·62197)

The values adopted for all parameters are those recommended by the Aerology Commission of O.M.I. (Publication No. 79, 1951). In this same publication will be found tables of values for e_w, e_i, L_f, L_s and L_v.

3. The water-saturation pseudo-adiabat

The equations describing the reversible and pseudo-adiabatic expansion of air saturated with respect to a plane surface of liquid water are

$$\left.\begin{aligned} \int (C_p + r_w C)\, d(\ln T) + \frac{r_w L_v}{T} - R \ln p_a &= A_1 \\ r_{cw} &\equiv 0 \end{aligned}\right\} \quad . \quad . \quad (1)$$

where A_1 is a constant.

The equations have the property that once initial pressure and temperature value are prescribed the subsequent states of the air are uniquely determined. The equation of the water-saturation adiabat (Eq. 3) require, in addition, that the total mixing rati of water in both vapour and liquid form be prescribed. Expressed geometrically, throug each point of a p-T plane there passes one water-saturation pseudo-adiabat but an infinit of water-saturation adiabats. The statement holds, too, for temperatures $T \leqslant 0$ °C i 'ice' be substituted for 'water.'

For this reason of uniqueness, the pseudo-adiabatic relation has become almos universally adopted to describe the behaviour of expanding saturated air. Thus th familiar *saturation adiabats* of aerological diagrams trace the pressure-temperature relation ship of the pseudo-adiabatic process.

Recently, a table of the numerical values of the water-saturation pseudo-adiabati process* has been prepared by Godson (1953) using modern and approved data. Sinc aerological investigations at Imperial College are carried out on Tephigrams inscribe with isopleths of this process (Imperial College Tephigram and British Meteorologica Office Tephigram Form 2810A, 1955), I have adopted it as the *standard*, with which al comparisons are made.

4. The ice-saturation pseudo-adiabat

The equations describing the reversible and pseudo-adiabatic expansion of air saturated with respect to a plane surface of ice are

$$\left.\begin{aligned} \int (C_p + r_i\, C_i)\, d(\ln T) + \frac{r_i\, L_s}{T} - R \ln p_a &= A_2 \\ r_{ci} &\equiv 0 \end{aligned}\right\} \quad . \quad . \quad (2)$$

where A_2 is a constant. (The process is, of course, defined only for temperatures $T \leqslant 0$°C).

In the absence of numerical values for this process comparable with those computed by Godson, and described in Section 3, I have prepared a skeleton table of values for the temperature range 0°C to − 50°C. (A method similar to Godson's was used).

If, now, the temperature attained in ice-saturation pseudo-adiabatic expansion from an initial state with temperature 0°C be compared with that attained in water-saturation pseudo-adiabatic expansion from the same initial state, it is found that, for the same pressure, the former exceeds the latter. Isopleths of the difference, which arise from the greater latent heat released by the condensation of vapour into ice, are shown in Fig. 1.

(The coordinates of this diagram require comment. The abscissa of potential temperature is the temperature which air undergoing a given *water-saturation* pseudo-adiabatic process attains at a pressure of 1,000 mb; the ordinate is the temperature attained in this process).

The difference steadily increases from zero at 0°C to a value of about 0·8°C at − 10°C, reaches a maximum at about − 30°C and finally slowly decreases. The decrease at low temperatures is due to the fact that the quantity of vapour available there for condensation is smaller in the 'ice' process than in the 'water' process. The maximum difference ranges from 1·2°C to 2·4°C.

A recommendation that these differences appear on aerological diagrams was adopted by O.M.I. in 1947, but they have not been published before.

* Even at sub-zero temperatures saturation is with respect to liquid water.

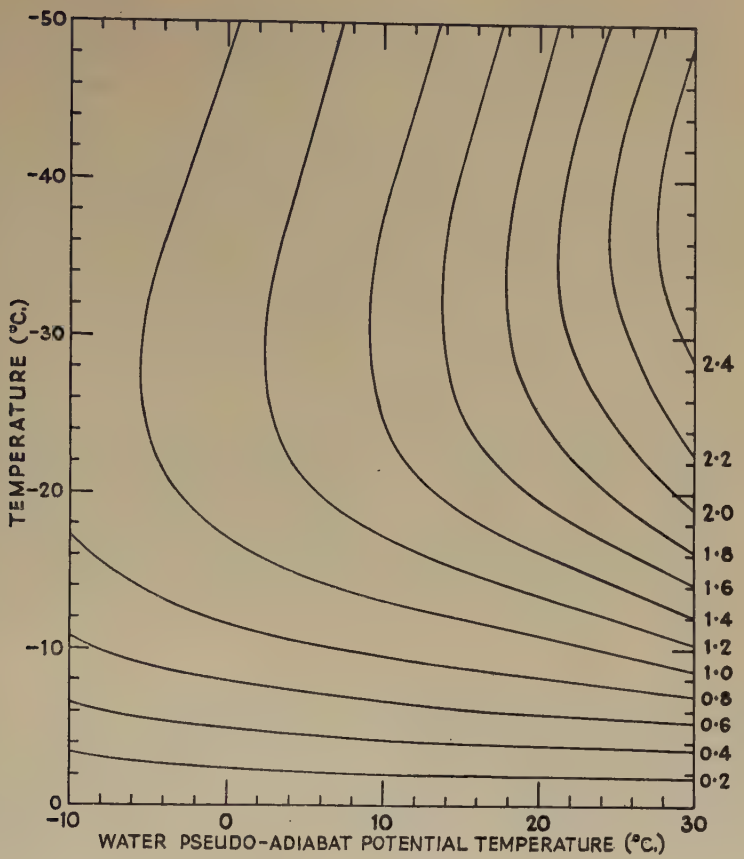

Figure 1. The difference in °C between the temperature attained in ice-saturation pseudo-adiabatic expansion and that attained in water-saturation pseudo-adiabatic expansion from the same initial state at 0°C (the former exceeds the latter).

5. The water-saturation adiabat

The equations describing the reversible and adiabatic expansion of air saturated with respect to a plane surface of liquid water and containing a liquid-water condensate are

$$\left.\begin{aligned}(C_p + rC)\ln T + \frac{r_w L_v}{T} - R\ln p_a &= A_3\\ r_{cw} &= r - r_w(T)\end{aligned}\right\} \quad . \quad . \quad (3)$$

where A_3 is a constant and where the total mixing ratio of water, r, is invariant. It may be noted that in its derivation it is assumed that the perfect gas law is obeyed; that is, the fraction of the air volume occupied by the liquid water is supposed negligible. This relation, of course, describes the expansion of air when its temperature falls below 0°C providing the condensate remains liquid.

I have computed the difference, at the same pressure, between the temperature attained in adiabatic expansion (Eq. 3) and that attained in pseudo-adiabatic expansion (Eq. 1) from the same initial state. Isopleths of this difference are shown in Figs. 2 and 4. An initial value of 900 mb is taken for the pressure in Fig. 2 and of 800 mb in Fig. 4; a range of initial temperatures is chosen which is meteorologically appropriate. The

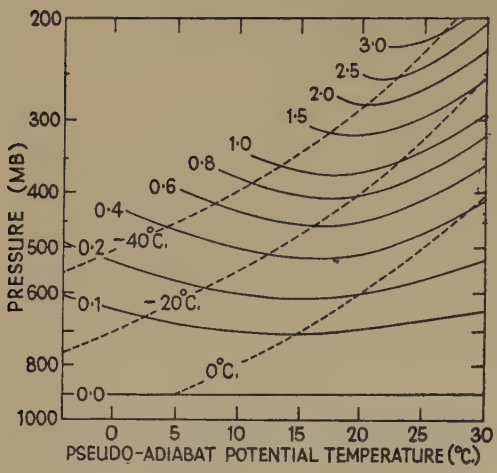

Figure 2. The difference in °C between the temperature attained in water-saturation adiabatic expansion and that attained in water-saturation pseudo-adiabatic expansion from the same initial state at 900 mb (the former exceeds the latter).

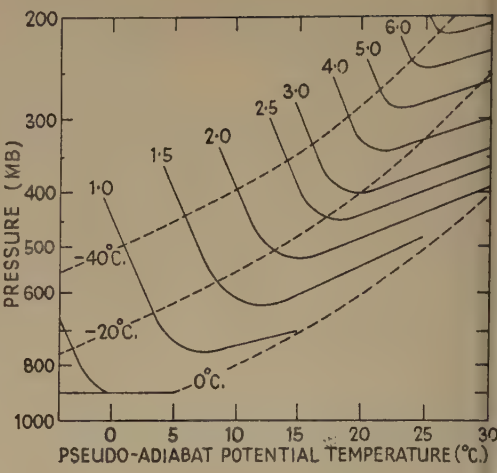

Figure 3. The difference in °C between th temperature attained in ice-saturation adiabat expansion and that attained in water-saturatio pseudo-adiabatic expansion from the same initi state at 900 mb (the former exceeds the latter

total mixing ratio of water in air expanding adiabatically, the r of Eq. 3, is taken as th saturation-mixing ratio with respect to liquid water *at* the initial pressure and temperature that is in its initial state the liquid-water content of air expanding adiabatically is zero.

Air which is saturated with respect to liquid water and which expands *adiabatically* is warmer at a given pressure than that which expands *pseudo-adiabatically* (given th same initial state). For any change of pressure, corresponding to a given work of expansion the air of the adiabatic process may draw on a greater reservoir of internal energy becaus of the liquid water present; consequently it cools less. For an initial pressure of 900 mb the difference is approximately $\frac{1}{2}$°C after 400 mb pressure drop and exceeds 1°C onl after 550 mb. For initial states of lower pressure the differences are smaller as the amoun of liquid water involved becomes less. (The meteorological significance of these difference can be appreciated if it is recalled that in the examination of certain aspects of clou behaviour the temperature of the cloud air needs to be specified with an accuracy o about 0·2°C).

6. The ice-saturation adiabat

The equations describing the reversible and adiabatic expansion of air saturate with respect to a plane surface of ice and containing an ice condensate are

$$\left.\begin{aligned}(C_p + rC_i)\ln T + \frac{r_i L_s}{T} - R\ln p_a &= A_4\\ r_{ci} &= r - r_i\,(T)\end{aligned}\right\} \quad . \quad . \quad (4$$

where A_4 is a constant and where the total mixing ratio of water, r, is again invariant

It is appropriate to recall that the majority of tropospheric clouds form initially a clouds of water droplets; that is, the majority of ice clouds are transformed from wate clouds. In the following paragraphs I have distinguished between expansions of clou air in which the conversion from water to ice takes place at a temperature of 0°C, an expansions in which the conversion takes place only after supercooling.

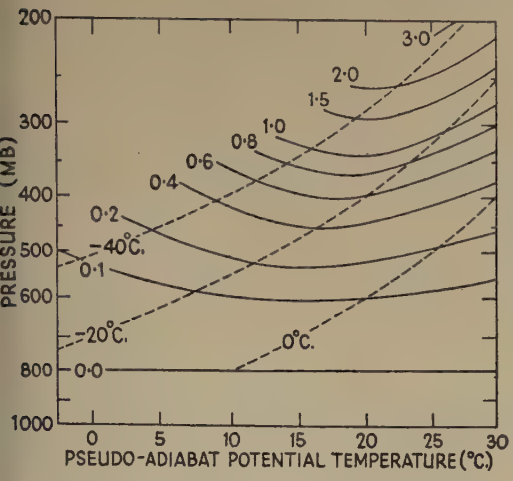

Figure 4. As Fig. 2 but for an initial pressure of 800 mb.

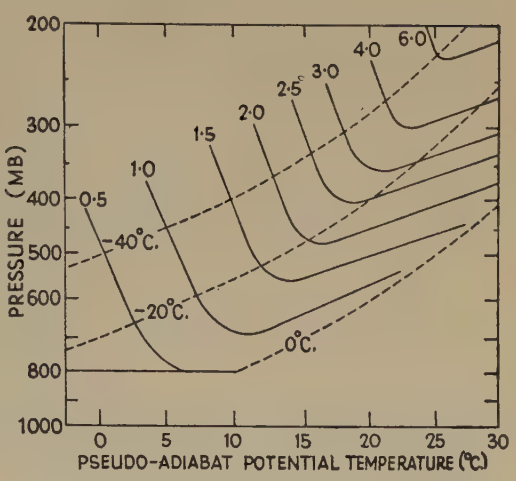

Figure 5. As Fig. 3 but for an initial pressure of 800 mb.

(*a*) *Isothermal conversion at* 0°C

When the freezing of liquid water commences at 0°C, (latent) heat energy is made available for expansion of the air, the temperature remaining at 0°C until all the liquid water is converted to ice. Graphically, this stage is represented by a *step* in the curve describing the adiabatic process. At temperatures $T > 0$°C the behaviour of the expanding cloud is described by Eq. 3 and for temperatures $T < 0$°C by Eq. 4. Moreover the A_4 and A_3 of these equations are related by the further equation

$$A_4 = A_3 - r\,(C - C_i)\left\{\ln T_0 - \frac{L_f(T_0)}{T_0}\right\} \quad . \quad . \quad . \quad (5)$$

where T_0 is the freezing temperature on the absolute scale. (This latter result may be derived from Eqs. 7, 9, 11, of § 34, 35 in Chapt. 3 of Brunt 1952).

Using Eqs. 3, 4, 5 of this discussion Figs. 3 and 5 have been prepared. They show the difference between the temperature attained in the ice-saturation *adiabatic* process and that attained in the water-saturation *pseudo-adiabatic* process. The initial pressures of Figs. 3 and 5 are 900 mb and 800 mb respectively, corresponding to Figs. 2 and 4.

The temperature differences are in part due to the presence of the condensate (as described in a previous section) and in part due to the latent heat released by the freezing of the condensate and by condensation of the vapour to ice. It is evident that these differences may be considerable; even with a (water saturation) pseudo-adiabat potential temperature of only 5°C they may amount to 1°C, and with potential temperatures of 25°C they may amount to 5°C.

(*b*) *Isobaric freezing at sub-zero temperatures*

Consider now the behaviour of a liquid-water cloud cooled by expansion to a sub-zero temperature at which freezing commences. Such a process is spontaneous and therefore irreversible, but a simple treatment can be given if the conversion takes place without change of pressure.

In an adiabatic and isobaric process enthalpy is conserved; from this fact we may deduce the temperature T' resulting from the freezing of a supercooled cloud of droplets initially at a temperature T to be given by

$$(C_p + rC_p')(T' - T) = \{r - r_w(T)\} \cdot L_f(T) + \{r_w(T) - r_i(T')\} \cdot L_s$$

The equation, which has an obvious physical interpretation, may be restated in the following approximate form suitable for computation

$$T' - T = \frac{(r - r_w) L_f + \frac{\epsilon L_s}{p'}(e_w - e_i)}{C_p + rC_p' + \frac{\epsilon L_s}{p'} \cdot \frac{de_i}{dT}} \quad . \quad . \quad . \quad (6)$$

where all the temperature-dependent terms have values appropriate to the temperature T, and p' is the pressure at which the freezing takes place. After complete conversion the values of T', p' and r when substituted in Eq. 4 determine the subsequent behaviour of the expanding air. Equivalently they define the value of A_4.

Extensive computations show that the value of A_4 is almost independent of the temperature of freezing. This is illustrated by the following alternative transformations of air saturated under a pressure of 700 mb with a liquid-water content of 3·0 g per Kg at a temperature of 0°C. (a) The liquid-water condensate freezes isothermally at 0°C; further expansion with an ice condensate and ice saturation results in a temperature of − 40°C at 323 mb. (b) The air expands with a supercooled condensate and with water saturation to − 20°C, there freezes isobarically and continues to expand with an ice condensate and ice saturation; a temperature of − 39·9°C is attained at 323 mb. (c) The air expands with a supercooled condensate and water saturation to − 40°C where it freezes isobarically; it attains, after further expansion, a temperature of − 39·7°C at 323 mb. In these adiabatic processes the state of ice-saturated air with an ice condensate is almost independent of its previous *particle* history.

The cause of the dependence, slight as it is, may be explained with the aid of the concept of entropy. The entropy of air, vapour and water is unchanged during its reversible and isothermal freezing at 0°C; at any lower temperature irreversible freezing results in an increase of entropy. Noting that the A_4 of Eq. 4 is a measure of the entropy of air, vapour and ice, it is readily demonstrated that an increase in A_4 results in an increase in temperature at any pressure. Accordingly we may state the following proposition. Given that a mass of saturated air undergoes only adiabatic and reversible changes of pressure, any temperature which it attains after the freezing of condensed water at a sub-zero temperature is higher than the temperature attained at the same pressure after freezing isothermally at 0°C. My computations show that within the domain of Figs. 3 and 5 the difference does not exceed 0·2°C if the freezing takes place at − 20°C, and does not exceed 0·5°C if the freezing takes place at − 40°C.

7. Cloud virtual temperature

To complete my discussion of adiabatic changes of state I shall briefly consider the effect of the weight of a dispersed condensate of water or ice on the density of saturated air.

The density of unsaturated air with vapour-mixing ratio r_v is given by the familiar expression

$$\rho = \frac{p}{RT_v} = \frac{p(1 + r_v)}{RT(1 + r_v/\epsilon)} \quad . \quad . \quad . \quad . \quad (7$$

where T_v is the virtual temperature, $T_v > T$. For saturated air the appropriate saturation mixing ratio is substituted for r_v.

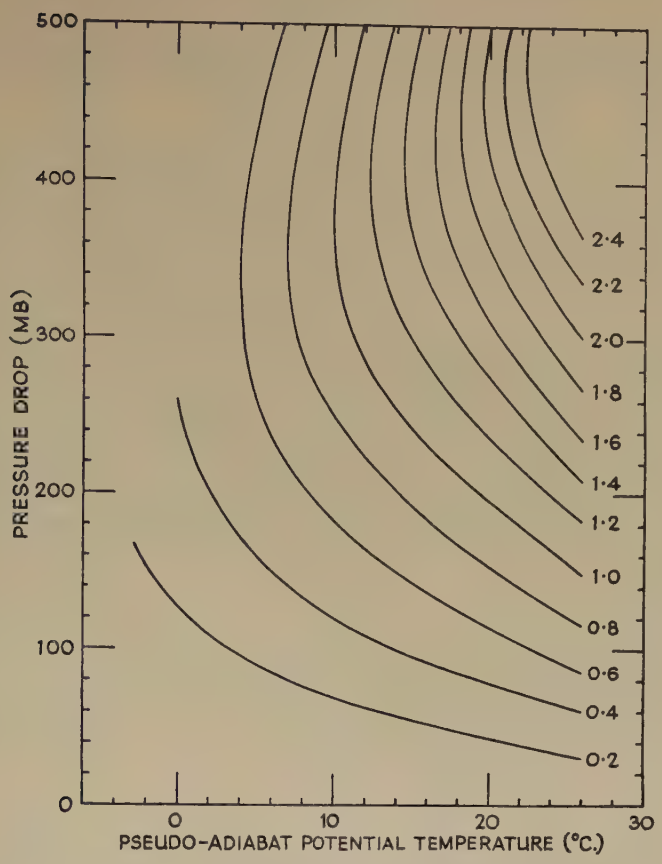

Figure 6. The difference in °C between the cloud virtual temperature attained in water-saturation adiabatic expansion and that attained in water-saturation pseudo-adiabatic expansion from the same initial state (the former is less than the latter).

In the presence of droplets with mixing ratio r_{cw}, the density as defined above is augmented by an amount

$$\Delta\rho = \rho \cdot r_{cw}$$

or equivalently the virtual temperature can be regarded as reduced by an amount

$$-\Delta T_v = T_v \cdot r_{cw} \quad . \quad . \quad . \quad . \quad . \quad (8)$$

The quantity $T_v + \Delta T_v$, for which the name *cloud virtual temperature* is suggested, then defines cloud density.

From Eqs. 3, 7, 8 the cloud virtual temperature for the water-saturation adiabatic process has been computed. Differences between the cloud virtual temperature for this process and that of the water-saturation pseudo-adiabat are shown in Fig. 6. (For the pseudo-adiabatic process $r_{cw} \equiv 0$ and cloud virtual temperature and virtual temperature are the same). The ordinate of the diagram, it may be noted, is the change in pressure from its initial value; thus the diagram may be used for a range of initial pressures. For a range 950 to 750 mb it will give an accuracy of 0·1°C for pseudo-adiabat potential temperatures less than 10°C and 0·2°C for potential temperatures greater than 20°C.

It is evident that, despite the temperature excess of the adiabatic process over that of the pseudo-adiabatic process (Figs. 2 and 4), because of the weight of the liquid water,

the cloud virtual temperature of the former is *less* than that of the latter. The difference may amount to 1°C with a pressure decrease of 200 mb and may rise to 2°C with a larger change.

8. Conclusions

Let us consider the more important implications of these investigations.

The difference between similar* reversible adiabatic and pseudo-adiabatic expansions is negligible, as is sometimes stated, *but only* when the change of pressure is small *and also only* when there is no change in the nature of the condensate.

Thus, in general, in order to determine cloud properties on the basis of adiabatic ascent, corrections must be made to those values deduced from, say, aerological diagrams which bear isopleths of a pseudo-adiabatic process.

Where our particular interest is with cloud virtual temperature, which effectively determines the *buoyancy* in cloud, the corrections of Fig. 6 must be applied, providing that the cloud contains only droplets. The observation that the formation of precipitation in middle-latitude cumulus is accompanied by the extensive appearance of the ice phase, commonly called glaciation (Ludlam and Saunders 1956), suggests that in so far as the adiabatic and reversible hypothesis is valid these corrections will only apply to a non-precipitating cumulus.

As we have seen, the freezing of the liquid-water content in adiabatic expansion results in a considerable increase in temperature and therefore in the buoyancy of cloud air. This mechanism may be invoked to interpret the behaviour of a cumulonimbus whose growth suddenly increases shortly after the first traces of glaciation in its summits, as will be discussed in a future paper.

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* Similar in respect of saturation.

551.46 : 532.49

Attenuation of swell in the North Atlantic Ocean

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(Manuscript received 18 January 1957, in revised form 30 April 1957)

Summary

Waves recorded in the decay area by a ship-borne wave recorder on o.w.s. *Weather Explorer* are compared with estimates of waves in the generating area. By restricting attention to the wave energy contained within one-second wave period intervals, it is possible to eliminate the effect of dispersion and to study the decrease in height due to loss of wave energy and the broadening of the wave front. While it is not possible to give an accurate assessment of the relative magnitude of these effects, the formula obtained in the case of calm conditions or following winds in the decay area is $H_T/H_T^2 = (300/R)^{\frac{1}{2}}$, where R is the distance of the recording point from the centre of the generating area. This formula does not depend on wave period and following wind speed, and suggests that the effect of broadening the wave front is the more important. It has been used successfully in predicting wave heights and periods at Casablanca.

1. Introduction

The work described in a previous paper (Darbyshire 1955), dealt with waves generated near the recording station, so that the characteristics of the waves and the wind could be related directly without allowance for attenuation and possible departures from the theoretical group velocity after the waves leave the generating area.

The present paper describes an investigation of waves after they have left the generating area and have travelled through the so-called decay area. The ideal requirement is for more than one recording point in the decay area as well as records of waves inside the generating area, but, in view of the practical difficulties, the work has to be based on records taken in the decay area by o.w.s. *Weather Explorer* and estimates of the waves in the generating area by formulae obtained in the previous paper. Twelve examples were selected in which the ship's position was 400 to 1,600 mi from the storm area, during the period February 1953 to December 1955.

The object of the first part of the investigation was to check the assumption made in my previous paper (1955), that there was no significant difference between the period of the storm waves and the resulting swell, so that the wave energy associated with a given wave period could always be identified and it would then be possible to study the loss of height as the swell travels from the generating area.

2. Analysis of wave spectra

As described previously (Darbyshire 1955), the wave records were converted into a black and white profile form for Fourier analysis by the wave analyser described by Barber, Ursell, Darbyshire and Tucker (1946). This analysis takes the form of a series of peaks each corresponding to periods which are sub-multiples of the duration of the record analysed. These spectra are reduced to a simpler form in the way previously described (Darbyshire 1955). Using the same notation, H_T is the square root of the sum of the squares of the peaks for each one-second period interval $(T - \frac{1}{2})$ to $(T + \frac{1}{2})$ sec, (where T is a whole number), and is the height of a simple sine wave having the same energy as the total energy of waves in that period interval. Similarly, taking the whole spectrum, H (the equivalent wave height) is equal to the square root of the sum of the squares of the heights

of all the peaks in the spectrum, or the height of a simple sine wave having the same energy as the total wave energy. It follows that

$$H^2 = \sum_0^\infty H_T^2$$

It was also shown (Darbyshire 1952) that the maximum height on a record of about 100 waves, H_{max}, is equal to $2H$. As the wave records were of rather short duration, about 10 to 12 min, the values of H_T obtained are subject to random variation (Tucker 1957), but this can be allowed for to some extent, as in the previous paper, by taking a running mean across the spectrum so that

$$H_T' = \tfrac{1}{4}(H_{T-1} + 2H_T + H_{T+1})$$

As $\Sigma H'^2_T \neq \Sigma H_T^2 (= H^2)$, it is not possible to evaluate H from the smoothed values but as the value of H is not used in this part of the investigation this is no great disadvantage.

To simplify the investigation, attention was confined to cases where the swell could be traced to one storm area, and 12 were found, involving a total of 100 wave records between Feburary 1953 and December 1955. The storms varied from 400 to 1,000 mi in extent. The H_T values for each example were plotted against the time and Fig. 1 shows such a series for waves recorded at station 'Kilo' (45°00'N, 16°00'W), on 2 to 3 September 1955. These waves could be traced back to a tropical storm 'Edith' which passed over the Bermuda area on 30 to 31 August 1955. Fig. 2 shows a similar example

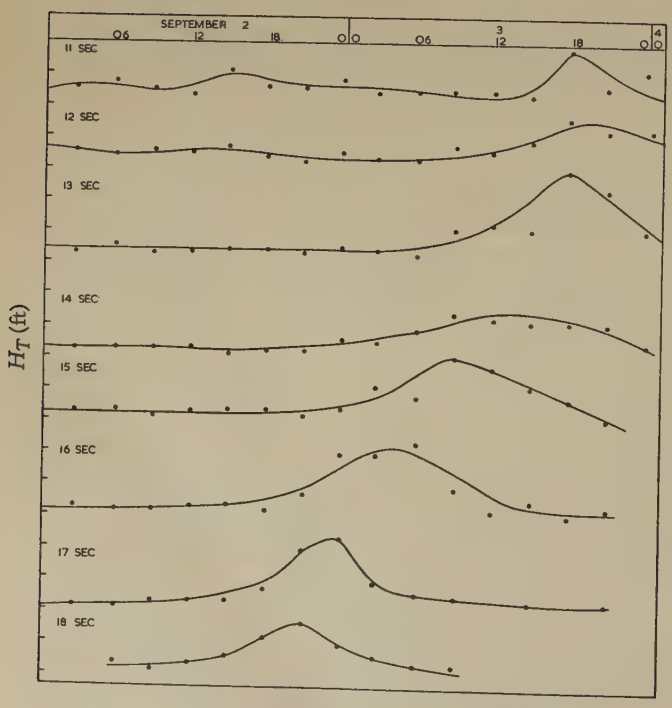

Figure 1. Plot of H_T against time, 2-3 September 1955.

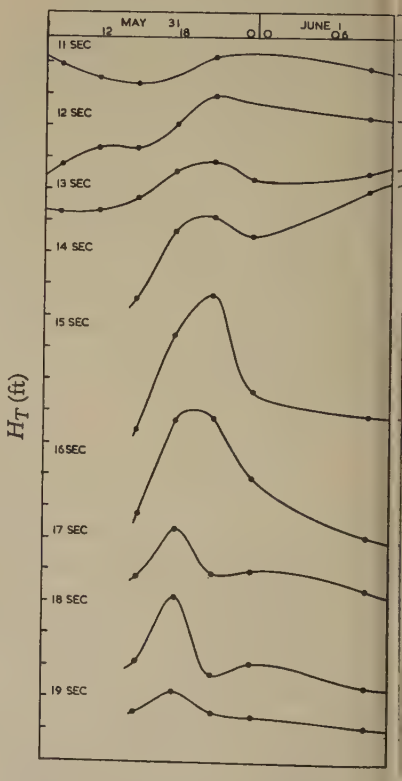

Figure 2. Plot of H_T against time, 31 May to 1 June 1954.

The waves were recorded at station ' Juliett ' (52°20'N, 20°00'W), on 31 May to 1 June 1954; they were traced to a storm situated about 600 mi west of the recording station. Similar diagrams were drawn for the other storms considered. The wind data were obtained from the six-hourly charts supplied by the Naval Weather Service, and were plotted at different times and distances from the ship on a ' wind data diagram ' similar to that described previously (Darbyshire 1952) using the ship's position as the reference point. Wind speeds cannot be charted over the sea as accurately as on land because there are fewer observations and this must introduce one element of uncertainty into the investigation.

3. Period of swell

The H_T values for each wave period tend, as indicated in Figs. 1 and 2, to reach a maximum value at a particular time, and if the waves arriving at these times are traced back by using wave-propagation lines, whose slopes correspond to the theoretical group velocity (Darbyshire 1952) they generally pass through the strongest part of the storm. From examination of the H_T diagrams for each storm, the maximum period with a measurable H_T value, T_m, and the wave period with the highest H_T value, T_s, could be determined. These two values were compared with the corresponding values for storm waves obtained in the previous paper (1955). The comparisons are shown in Figs. 3 (a) and 3 (b), where the vertical scales represent T_m and T_s, and the horizontal scale, U, the highest value of gradient wind speed which has persisted over more than 200 mi in the storm area. The distance 200 mi is used because a fetch longer than this is believed not to increase the wave energy appreciably (Darbyshire 1955, 1956). However, in the examples

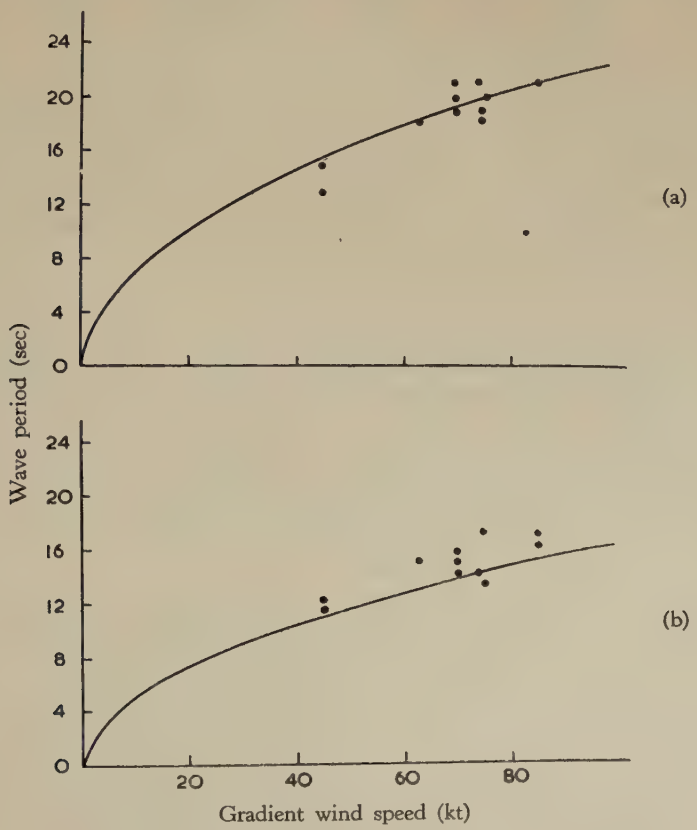

Figure 3. Variation of T_m and T_s with gradient wind speed.

used, the wind speed was constant within about 5 kt for distances varying from 200 to 700 mi, the mean being 400 mi. The full lines in Figs. 3 represent the relations found in the previous paper.

$$T_m = 2{\cdot}30\ U^{\frac{1}{2}}\ (U, \text{kt};\ T, \text{sec}) \quad . \quad . \quad . \quad . \quad . \quad (1)$$

$$T_s = 1{\cdot}64\ U^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad . \quad . \quad (2)$$

The points lie fairly symmetrically about the curves obtained for storm waves (1955), with about the same scatter. Although the comparison is indirect, it suggests that there is no appreciable interchange of energy between waves of different period after leaving the generating area. It will therefore be assumed that the H_T value corresponding to a period T in the decay area can be identified with the H_T value for the same period T in the generating area, H_T°.

4. Variation of H_T/H_T° from the centre of the generating area

To evaluate H_T°, use is made of the equation

$$H_T^\circ = 0{\cdot}0036\ (U^{\frac{1}{2}} - 0{\cdot}43\ T)\ T^{\frac{5}{2}} \quad . \quad . \quad . \quad . \quad (3)$$

$$H_T^\circ = 0 \text{ for } U^{\frac{1}{2}} < 0{\cdot}43\ T\ (H_T^\circ, \text{ft};\ U, \text{kt},\ T, \text{sec})$$

which was obtained in the previous paper (1955). This equation was based on points involving a good deal of scatter, most of which could be attributed to random variation (Tucker 1957), already referred to, and to errors in estimating the wind speed. The effect of wind-speed errors can be indicated by differentiating Eq. (3) and dividing, when one obtains

$$dH_T^\circ/H_T^\circ = 0{\cdot}5\ (dU/U)\ (1 - 0{\cdot}43\ T/U^{\frac{1}{2}})^{-1}$$

hence for a given value of T, errors in H_T° due to errors in U will decrease with wind speed. It also follows from Eq. (3) that the greater the wind speed, the greater the value of H_T° for a given wave-period T so that the error due to the wind speeds for a given wave period is least for the maximum value of H_T°. Accordingly, H_T diagrams like those shown in Figs. 1 and 2 were used to pick out the largest H_T value for each wave period from each storm. Then to estimate H_T°, waves of period T at the time of the maximum H_T value are traced back to the storm area, using the wave propagation lines, the highest value of the gradient wind speed that had acted on the waves for a distance of more than 200 mi (see above) being found and substituted in Eq. (3). The ratio between the observed maximum values of H_T and the corresponding computed values of H_T° were then plotted against the distance from the storm centre in Fig. 4 for periods ranging from 13 to 18 sec. It is not possible to use each period for every storm since, in some, the higher periods were not generated and in others the energy of the lower-period waves was obviously supplemented from other generating areas. There appears to be no systematic variation of H_T/H_T° with period. The curve in each diagram corresponds to

$$H_T/H_T^\circ = (300/R)^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

where R is the distance from the recording point to the storm centre. There were always calm conditions or following winds in the decay area, the mean gradient wind speeds in the decay area for the 12 examples being 26, 25, 20, 10, 0 (under 10) 0, 10, 10, 40, 33, 20, and 10 kt. These figures are noted in the diagrams in Fig. 4 and so far as can be judged, the ratio H_T/H_T° is not dependent on the speed of following winds in the decay area. The effect might well be different for opposing winds and possibly also for cross winds but those conditions are difficult to study because the weather situations required are rarely found in the eastern half of the North Atlantic.

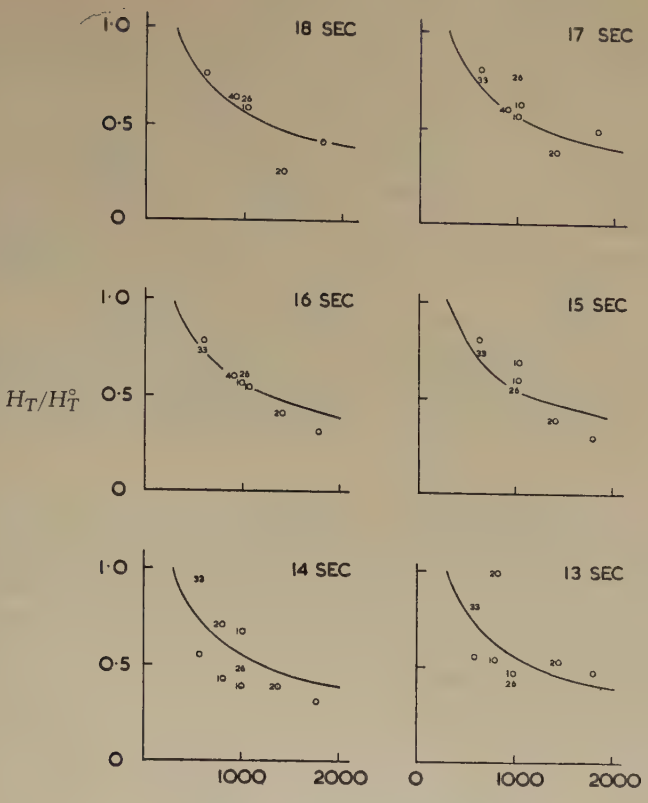

Figure 4. Variation of H_T/H_T° with distance from storm centre.

5. Possible causes of the variation of H_T/H_T° with distance

The variation of the ratio with distance could be due to three effects.

(*a*) *Dispersion*

Waves of different periods have different group velocities so that a certain amount of wave energy, spread over a wide range of periods and confined initially to a given area, will after a lapse of time be distributed over a larger area and the energy density will be reduced. This effect is not appreciable if attention is restricted to a narrow period range $(T - \frac{1}{2})$ to $(T + \frac{1}{2})$ sec and the winds in the storm area do not change materially during the three hours or so which elapse between the times at which the slower waves of $(T - \frac{1}{2})$ sec period and the faster waves of $(T + \frac{1}{2})$ sec must depart from the storm area to arrive at the recording point simultaneously. This condition seems to be fulfilled for all the examples used for T from 13 to 18 sec.

(*b*) *Loss of wave energy*

We still do not know how waves gain and lose their energy and since the theories put forward so far are of tentative character, it may be useful to review them in the light of the results shown in Fig. 4.

One suggested cause of attenuation is the effect of air resistance, discussed by Sverdrup and Munk (1947) as a development of the theory of wave generation given by Jeffreys

(1925, 1926). On this basis, the formula for attenuation becomes, when the surface wir speed W is in the same direction and is less than the wave speed c,

$$a/a_0 = \exp - \{gs\rho' (c - U)^2 x\}/2\rho c^4$$

where s is a non-dimensional constant taken by Sverdrup and Munk to be 0·12 an originally by Jeffreys to be 0·27. ρ' is the density of air, ρ, the density of water.

Another effect is that of turbulent friction which is usually represented by an edd viscosity ν_ϵ. It enters the equations in the same way and the formula for wave attenuatio on this basis is

$$a/a_0 = \exp - (4\nu_\epsilon g^2 x)/c^5$$

Values of ν_ϵ have been obtained mainly from measurements of sea currents caused by th wind. One such value, due to Thorade, is quoted by Sverdrup, Johnson, and Flemin (1946)

$$\nu_\epsilon = 1{\cdot}02\ W^3$$

where W is the surface-wind speed in m/sec and ν_ϵ is in CGS units. With this value 12 sec waves travelling through a region where the wind is 10 kt would be reduced t 1/10 in 500 mi. This effect is much too large and many writers have suggested that othe values for ν_ϵ should be applied to waves. Some have suggested that ν_ϵ depends on th actual wave characteristics, Bowden (1950) suggested it was proportional to ac, whil Groen and Dorrenstein (1950) and Groen (1954) suggested that $\nu_\epsilon \propto \lambda^{4/3}$.

All these formulae make H_T/H_T° depend on the wave period for a given value of th decay distance. The value of the wind speed in the decay area would also be importan if air resistance is the controlling factor and probably also if turbulent friction i considerable. These effects are not observed in Fig. 4, which, in spite of the scatter, give a fairly clear indication that the decrease of energy is independent of wave period an wind speed.

(c) *Divergence*

Very little is known at present about the effect of divergence, or broadening of the wave front, when the waves leave the generating area. Eq. (4) corresponds, however, to a simple inverse square-root law of attenuation which applies to the simplest form of divergence, where the waves are non-dispersive and the wave fronts are always contained within a sector centred on the storm area. This picture is probably an over-simplification of the actual conditions but an approximation to this state of affairs would be obtained if the two-dimensional spectrum formula was of a certain form. In general, the energy density per unit period, per unit angle, for a wind speed U is given by

$$E(T, U, \theta)$$

where θ is the angle the wave direction makes with the wind direction. If the variation of wave energy with direction is assumed to be independent of the wave period and wind speed, then it follows that

$$E(T, U, \theta) = f(T, U,)\, g(\theta)$$

If, furthermore, $g(\theta)$ is a function which is almost constant for a range of values centred round $\theta = 0$, and then falls off rapidly, it can be represented approximately by

$$g(\theta) = K \text{ for } -\theta_0 < \theta < \theta_0$$

$$g(\theta) = 0 \text{ for } \theta < -\theta_0 \text{ and } \theta > \theta_0$$

Under these conditions, the energy contained within an infinitesimal part of the generating area will diverge from it along a sector of angle $2\theta_0$ centred on it, and the attenuation will be of the form of Eq. (4). The effect is more complicated when it is integrated over all the generating area but would approximate to Eq. (4) when the distance from the storm area increases.

On the whole then, Eq. (4) corresponds more to the effect of divergence than any other suggested cause but in view of the indirect nature of the evidence and the scatter of the points in Fig. 4, a more exact assessment of the relative importance of the effect of divergence and that of real loss of energy must await measurements of wave spectra by two or more ships at different distances from the storm. Eq. (4) should at any rate provide a satisfactory empirical rule governing the rate of decay of swell under conditions in the North Atlantic Ocean where there are no opposing or cross winds, and in the next section it will be used in conjunction with Eq. (3) to forecast wave heights.

5. Application to swell at Casablanca

The wave heights and periods at Casablanca were predicted, as no other ship measurements were available, and it was considered that the measurements at this port would provide the best approximation to deep sea measurements. The coast, because of its unbroken nature, is well exposed to Atlantic swell. Strong tidal streams are absent and the continental shelf is very narrow, the 100-fathom line being about 40 mi from the

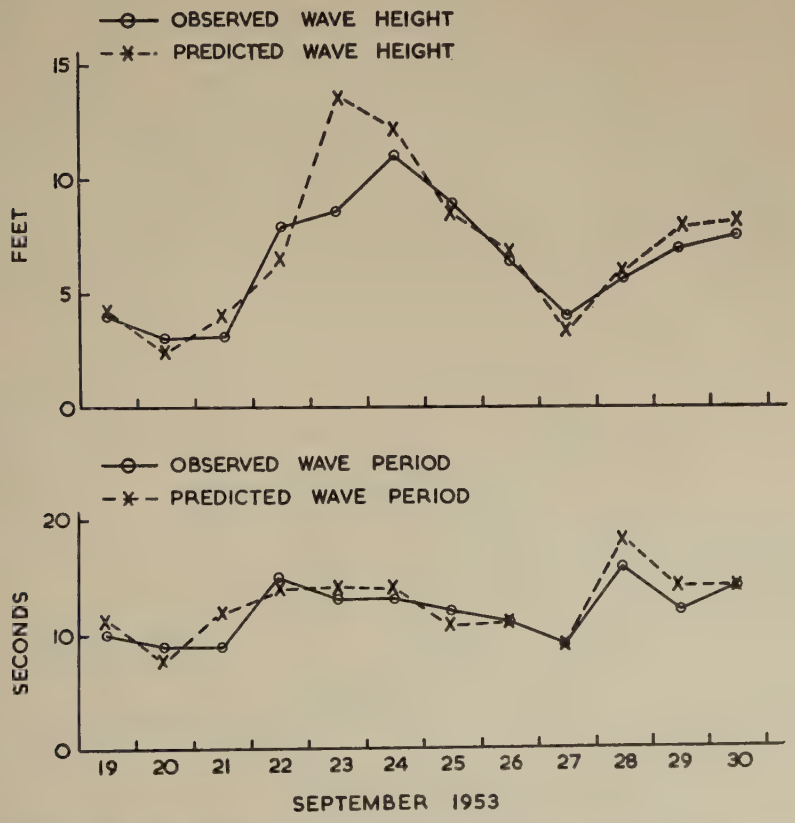

Figure 5. Plot against time of observed and predicted values of wave height and wave period, for Casablanca, September 1953.

coast. There is the possibliity of the waves being modified at the relatively shallow dept of recording, 20 m, but the effect on wave heights and periods should be small accordin to hydrodynamical theory. The periods should be unchanged and the heights are nc affected by more than 10 per cent. The effect of bottom friction is more difficult to asses but the narrowness of the continental shelf must reduce this effect considerably.

Wave measurements have been taken daily at 1100 hr since 1951 by a wave-pressur recorder laid at a depth of 20 m. According to Gelci (personal communication), the wav records are taken for about 20 min and the figures given for the height and perio correspond to the mean height of the highest tenth of the waves and the period of maximum amplitude on the wave spectrum.

Wind-data diagrams (Darbyshire 1952) were prepared from weather charts provided by the Naval Weather Service and they were used with wave-propagation lines based on the classical value of the wave-group velocity. H_T values for various wave periods were worked out for the times in question, using Eqs .(3) and (4). The period having the maximum H_T value was taken to be the most prominent period, and $2\,(\Sigma\, H_T{}^2)^{\frac{1}{2}}$ as the wave height since it is within 10 per cent of the mean of the highest tenth, (Longuet-Higgins 1952). Predictions were made for two continuous periods, 19 to 30 September 1953 and 3 to 28 February 1954. The storms producing the waves during these times were from 400 to 2,000 mi from Casablanca, and were similar in extent and magnitude to those from which Eq. (4) was obtained but the winds in the decay area were in all cases less than 10 kt. There were 38 predictions of wave height and period. These are compared with observed values in Figs. 5 and 6. Table 1 gives an analysis of the accuracy of the predictions.

The predictions of wave period are better than those of wave height but this is to be expected as the period depends on the square root of the wind speed and the height on square. Three-quarters of the wave-height predictions are, however, within 25 per cent of the measured values which is quite satisfactory when all the uncertainties are borne in mind.

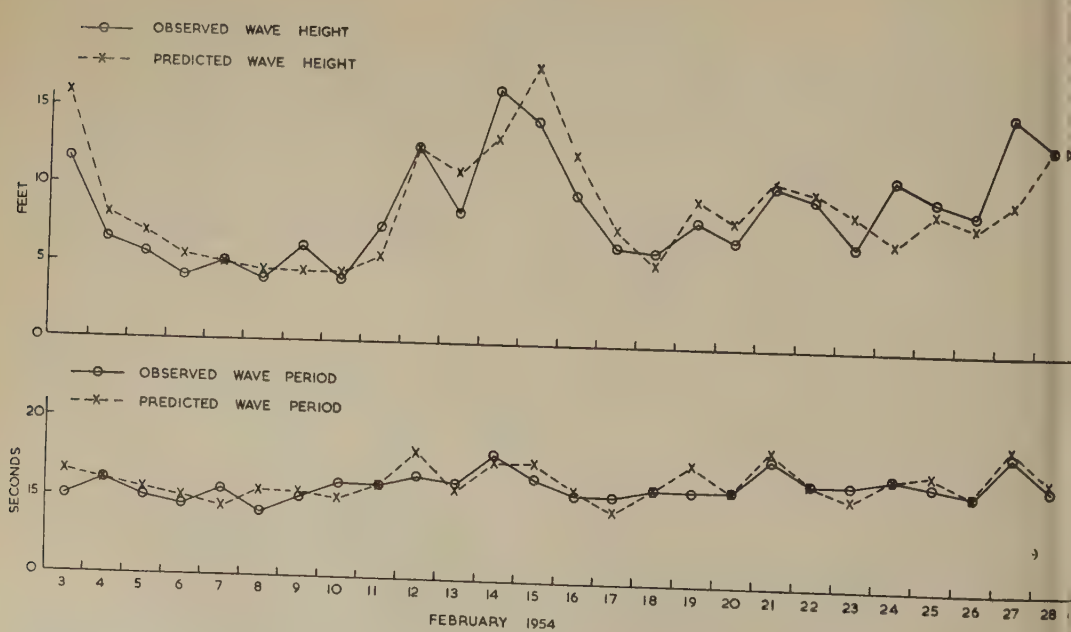

Figure 6. Plot against time of observed and predicted values of wave height and wave period, for Casablanca, February 1954.

TABLE 1. ACCURACY OF PREDICTIONS

Per cent error	Percentage of height predictions within per cent error	Percentage of period predictions within per cent error
5	21	37
10	34	66
15	47	77
20	63	90
25	74	92
30	82	97
35	87	100
40	97	

6. CONCLUSIONS

The periods associated with swell waves bear the same relation to the wind speeds in the generating area as do the waves in the storm. This makes it possible to identify the energy contained by waves of particular periods in the decay area with that contained by waves of the same period in the storm area.

This identification makes it possible to estimate the change in height of waves of a given period in the decay area, and the formula obtained, $H_T/H_T^\circ = (300/R)^{\frac{1}{2}}$, suggests that the broadening of the wave front is an important factor in influencing the loss of height of swell and the use of this formula for practical wave forecasting appears to be satisfactory.

ACKNOWLEDGMENTS

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Harmonic analysis of the mean northern-hemisphere wind field for the year 1950

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(Manuscript received 30 April 1957)

SUMMARY

Harmonic analyses of the mean northern-hemisphere wind field for the year 1950 along latitudes 30° 45° and 60° are presented. The seasonal variations of these spectra are investigated and the relative contributions of the various scales of mean eddies to the meridional transport of angular momentum are calculated.

1. INTRODUCTION

By examining the harmonic analysis of the mean-wind map it is possible to obtain valuable information concerning the nature of the standing eddies in the atmosphere. Of particular interest in this regard are,

(1) the determination of the wave lengths (or wave numbers) of the dominant eddies which are present on the mean map as a function of latitude,

(2) the determination of the seasonal variation of the amplitude and phase of the harmonics. This is of special interest in view of the current controversy regarding the relative importance of the heating and of mountains in giving rise to the standing eddies, and

(3) the determination of the contributions of the various scales of standing eddies to the meridional transport of angular momentum.

With a view to obtaining insight into these questions, harmonic analyses of the mean conditions for the year 1950, based on the 500-mb data presented by Buch (1954), have been performed. As a preliminary to presenting the results we shall briefly set down the basic relationships used in the study.

2. BASIC CONCEPTS

Any regularly-behaving function of longitude, $f(\lambda)$, specified along a given latitude, ϕ, and pressure surface, p, may be expressed in terms of a Fourier representation having the form,

$$f(\lambda) = [f] + \sum_{n=1}^{\infty} |F(n)| \cos n(\lambda - \epsilon(n)) \quad . \quad . \quad . \quad (1)$$

where the bracket, defined by

$$[(\,)] = \frac{1}{2\pi} \int_0^{2\pi} (\,)\, d\lambda \quad . \quad . \quad . \quad . \quad . \quad (2)$$

denotes the zonal average, n denotes the wave number around the latitude circle, and $|F(n)|$ and $\epsilon(n)$ are the amplitude and phase angle respectively of the nth harmonic component. These latter quantities are given explicitly by the relations,

$$|F(n)| = [\mathscr{F}_1^2(n) + \mathscr{F}_2^2(n)]^{\frac{1}{2}} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

and

$$\epsilon(n) = \frac{1}{n} \tan^{-1} \left\{ \frac{\mathscr{F}_2(n)}{\mathscr{F}_1(n)} \right\}, \quad . \quad . \quad . \quad . \quad . \quad (4)$$

$\mathscr{F}_1(n)$ and $\mathscr{F}_2(n)$ being the real and imaginary parts of the complex Fourier coefficient,

$$F(n) = \frac{1}{\pi} \int_0^{2\pi} f(\lambda)\, e^{-in\lambda}\, d\lambda$$
$$= \mathscr{F}_1(n) - i\mathscr{F}_2(n) \quad . \quad . \quad . \quad . \quad . \quad (5)$$

In the present study we shall consider the Fourier analysis of the eastward component of the ean wind $\bar{u}$, and the northward component of the mean wind $\bar{v}$. We shall denote the complex ourier coefficients corresponding to these quantities by

$$U(n) = \mathscr{U}_1(n) - i\,\mathscr{U}_2(n)$$

nd

$$V(n) = \mathscr{V}_1(n) - i\,\mathscr{V}_2(n)$$

espectively.

The spectral function for the meridional eddy transport of relative angular momentum cross latitude ϕ per unit pressure difference, due to the standing eddies, is given by the relation cf., Van Isacker and Van Mieghem (1956))

$$T(n) = \frac{\pi a^2 \cos^2\phi}{g}\left[\mathscr{U}_1(n)\mathscr{V}_1(n) + \mathscr{U}_2(n)\mathscr{V}_2(n)\right], \quad . \quad . \quad . \quad (6)$$

here a is the radius of the earth and g is the acceleration of gravity.

In performing the harmonic analysis it is, of course, necessary to select a finite number of armonics as the basis for the Fourier representation. In the present case harmonics up to wave umber 12 were computed, based on information at 36 points around the hemisphere.

3. Results

As noted in the introduction, in this study we are concerned with the spectral properties of he mean 500-mb wind field for the year 1950. In particular, the Fourier resolutions along the 30°, 45° and 60° latitude circles were selected for investigation.

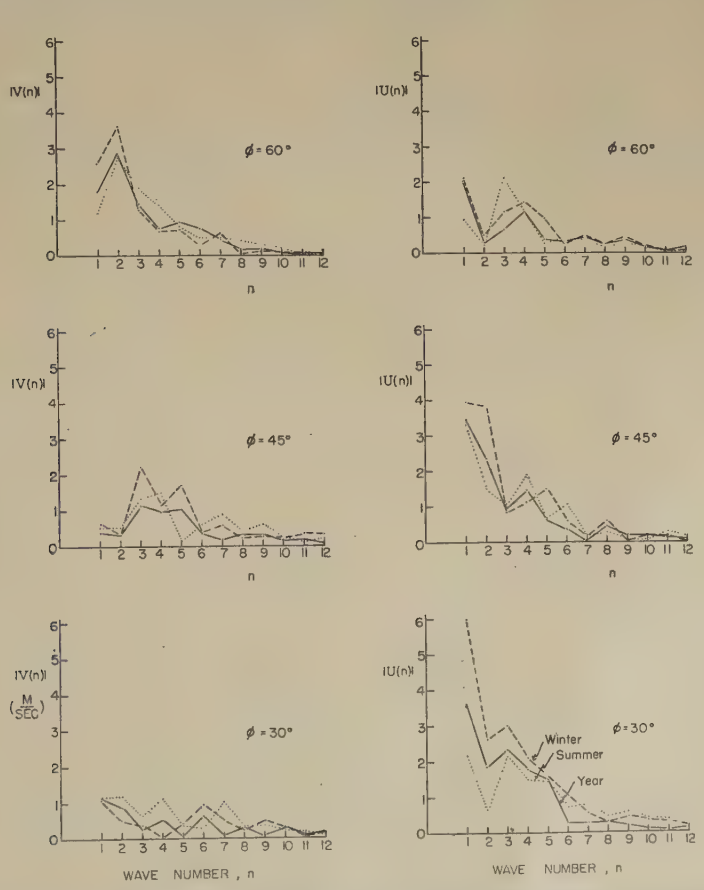

Figure 1. Amplitude spectra of meridional and zonal components of the 500-mb mean wind for the year 1950 denoted by $|V(n)|$ (first column) and $|U(n)|$ (second column), respectively. Solid line indicates *annual* mean; dashed line, *winter* mean; and dotted line, *summer* mean. Units are m/sec.

The mean-wind maps were constructed by Buch (1954) on the bases of average *actual* wind conditions reported at 81 stations in the northern hemisphere. ' Summer ' and ' winter ' means were obtained by dividing the year into two 6-month periods; (April-September) and (January-March, October-December) respectively. The spectra obtained for these seasons will therefore not necessarily agree with those obtained from the July and January mean maps to which reference is usually made (e.g., Scherhag 1948, Sutcliffe 1951).

The amplitude spectra, $|V|$ and $|U|$, are presented in Fig. 1. It should be noted that these are discrete ' line ' spectra and, as such, have meaning only for the integral values of wave number. Continuous lines have been drawn only to aid visually in tracing the variations of the spectra.

The variations of $\bar{v}$ around the hemisphere, measured in the wave number domain by $|V(n)|$ are closely indentified with the waves appearing in the mean streamline field. On the other hand the variations of $\bar{u}$, measured in the wave-number domain by $|U(n)|$, depend on the streakiness in the velocity of the mean zonal current around a latitude circle, associated, for example, with the meandering of the mean ' jetstream.'

The harmonics which are most prominent in the $|V|$-spectra are those of wave number 1 and 2 at 60°; wave number 3, 4 and 5 at 45°; and wave numbers 1, 2 and 6 at 30°. In general, the variations of $\bar{v}$ around the latitude circle are most pronounced at 60°, being of progressively smaller amplitude at 45° and 30°.

The amplitude spectra for the *zonal* component of the mean wind, $|U(n)|$, are shown in the second column of Fig. 1. It may be seen that most of the variation is concentrated in the longest waves with a maximum at wave number 1 at all three latitudes. This large amplitude of wave number 1 is a reflection of the fact that the mean hemispheric vortex is not symmetrical with respect to the pole, an occurrence which has been reported in detail by LaSeur (1954). The variations of zonal velocity are most pronounced at low latitudes, the reverse of the latitudinal distribution of $|V(n)|$. In general, the variation of the zonal component of the wind around a latitude circle is more pronounced than the variation of the meridional component, as evidenced by the higher amplitudes $|U(n)|$.

The spectral distributions for the winter and summer means are shown by the dashed and dotted curves respectively. For certain wave-lengths striking seasonal changes in the amplitudes are revealed. For example, $|V(5)|$ at $\phi = 45°$ is very small for the summer period whereas its value is quite high in the winter. Wave number 4 at $\phi = 30°$ shows a similar seasonal change with the maximum in this case occurring in the summer.

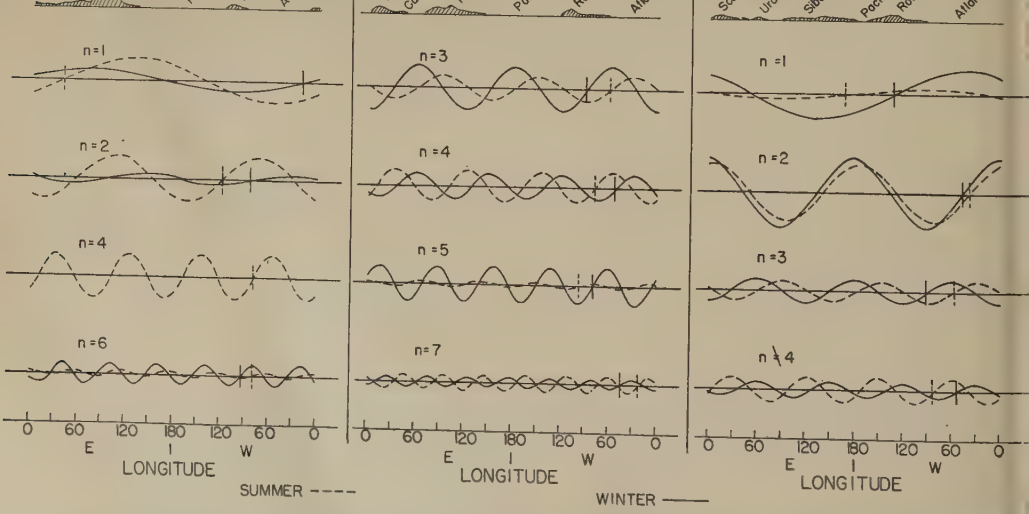

Figure 2. Schematic representation of the prominent waves which comprise the winter and summer mean fields of the meridional component of velocity. Trough-lines are located at the intersection of the abscissa and the wave progressing from a minimum to a maximum.

If the amplitude of an harmonic in the annual mean wind field lies nearly mid-way between he summer and winter values, it is implied that there is little phase shift of that Fourier component vith the seasons. It may be seen that in many cases this is not true, indicating that appreciable hase differences between harmonics in the winter and summer streamline patterns actually exist. n order to investigate this phenomenon in greater detail the phase angles ϵ (n) were computed for he four harmonics in the $\bar{v}$-field having the greatest amplitude at each latitude. The results are chematically represented in Fig. 2. The waves shown (drawn to scale) are those for the variations f the meridional component of the mean velocity, so that a ' trough line ' is located at the inter- ection of the abscissa and the curve which progresses from a minimum to a maximum. Short ertical lines have been drawn through this intersection to facilitate an inspection of the seasonal hase shift. In many cases these phase shifts are of the order of a quarter of a wave length, a otable exception being the high amplitude harmonic, $n = 2$, at 60°N which undergoes very ittle change in position.

In Fig. 3 the spectra of the meridional transport of angular momentum across the three atitudes studied due to the ' standing ' eddies are presented. The net transport at 30° and 45° s northward with a southward transport occurring at 60°. At all latitudes the longer waves are nost active in effecting the transport, wave number 1 being of particular importance during the winter at low latitudes.

4. Discussion

The observations reported here reveal some of the spectral properties of the mean hemispheric low pattern for the year 1950. These properties must be related to the irregularities in the surface of the earth manifested by differences in orography and heating. The relative importance of these

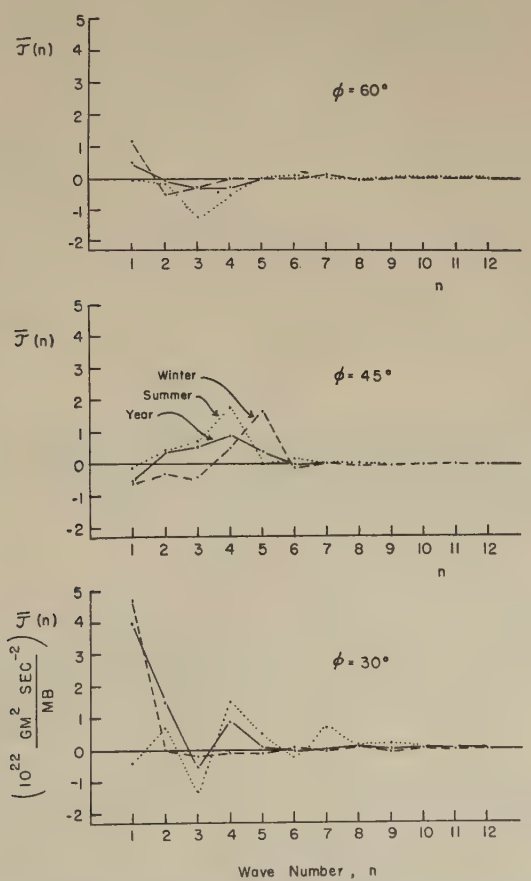

Figure 3. Spectra of the meridional transport of relative angular momentum per unit pressure difference across latitudes 30°, 45° and 60° due to the standing eddies. A positive sign indicates a northward momentum transport. Units are 10^{22} gm^2 sec^{-2} mb^{-1}.

two factors has been discussed by several writers (e.g., Charney and Eliassen 1949, Bolin 1950, Sutcliffe 1951, Smagorinsky 1953, Frenzen 1955) but, as yet, a complete theory by which one can explain observations of the type described here has not been achieved. In view of the complex nature of the spectral distribution of the mean flow and the irregular seasonal variation of the harmonics with regard to both amplitude and phase, it is likely that such a theory will involve both orography and heating in a complicated manner.

Acknowledgment

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551.521.63 : 551.510.534

The absorption coefficient of ozone at 3021 Å

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(Manuscript received 4 April 1957)

Summary

A new determination of the absolute value of the absorption coefficient of ozone at 3021 Å is described. The value obtained for the decimal absorption coefficient, measured at 20°C, is $3{\cdot}310 \pm 0{\cdot}009$ (cm STP)$^{-1}$. This figure is in excellent agreement with the value for this wavelength obtained by Vigroux (1953) and may be used with confidence in meteorological applications.

1. Introduction

Spectroscopic determinations of ozone amount or distribution in the atmosphere depend of the absolute values of the absorption coefficients of ozone in the ultra-violet. Most of such observations have been reduced using values determined by Ny and Choong (1933). Subsequent work and, in particular, a careful series of measurements by Vigroux (1953) have thrown considerable doubt on the accuracy of the earlier values, systematic discrepancies of the order of 15 per cent existing. While it is most important that all workers on atmospheric ozone should use the same mutually consistent values of absorption coefficients in the reduction of spectroscopic observations, it is also clearly desirable that these values should be as nearly as possible correct and a further independent check seems desirable.

This note describes a re-determination of the value of the absorption coefficient for the wavelength 3021 Å. This forms the first step in a programme aimed at establishing a series of standard absorption coefficients for ozone at definite and reproducible wavelengths.

* Now with International Ozone Commission, Oxford.

2. Method

The pressure of a mixture of ozone and oxygen in a sealed absorption tube is used as an indication of the amount of ozone remaining in the tube. This method was used in principle by Warburg (1902); we owe to Dr. R. M. Goody the suggestion that with modern techniques it may be made very accurate. Since two ozone molecules form, on decaying, three molecules of oxygen, then provided the experiment starts with nearly pure ozone in the tube, the pressure will increase by approximately 50 per cent as the ozone decays and gives, at any stage, an accurate indication of the decay without the necessity for any chemical analysis. A plot of optical density ($\log_{10} I_0/I$) against p should, if temperature changes are insignificant, as was the case in practice, yield a straight line whose slope is proportional to αl, where I_0 and I are intensities of the light entering and leaving the absorption tube respectively, p is the pressure, l the length of the column of gas and α the required decimal absorption coefficient.

Preliminary investigation showed that, with a pyrex absorption-tube, adsorption and desorption of ozone on the walls are of sufficient magnitude to make the method unreliable. With a quartz tube however this difficulty has been overcome.

Pure ozone is prepared, as described by Vigroux (1949), and distilled into the evacuated quartz absorption tube, approximately 10 cm in length, which is then sealed. The initial pressure is about 2 cm of mercury. The gas is then in contact only with quartz surfaces except for a small sensitive pyrex spiral pressure-gauge and one junction of a small 10 per cent Ir-Pt, 40 per cent Pd-Au thermocouple, sealed into the absorption tube by means of a thin-walled platinum tube. Light from a mercury lamp is passed through the absorption tube and the 3021 Å line is then isolated by a quartz Hilger spectrometer. The change in optical density of the absorption tube is measured at intervals as the ozone decays, by comparison with a reference beam, using a photomultiplier. Both beams come from the same small area of the source and fall on the same part of the photomultiplier cathode. The rate of the spontaneous decay, which is extremely slow in a quartz tube, is controlled by irradiation from a quartz mercury lamp.

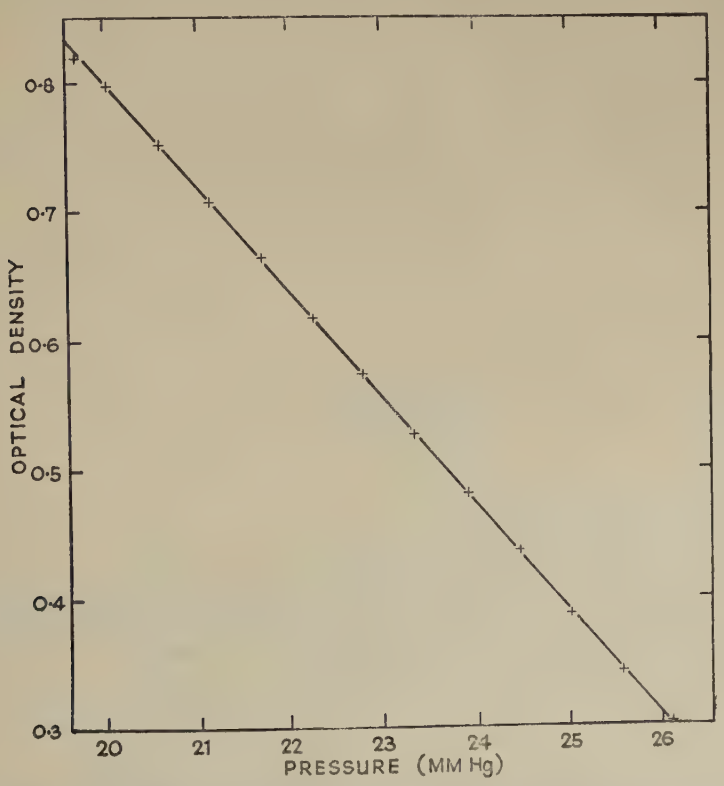

Figure 1. A typical plot of optical density ($\log_{10} I_0/I$) of the column of ozone against the pressure of the ozone-oxygen mixture, showing the linear relationship.

A plot of optical density against pressure normally yields a good straight line; Fig. 1 shows the results of a typical experiment. A small proportion of runs show an obvious kink or curvature in the relationship, apparently due to slight shifts in the apparatus; such runs are rejected. On a satisfactory run the slope of the line is determined with a standard error of the order of 0·3 per cent.

Impurities in the gas affect the determination only if they are unstable and cause pressure changes during the experiment; their effect has been shown to be quite negligible. Systematic errors caused by light scattered in the spectrometer were investigated by measuring transmissions at high optical density and were found to be less than 0·1 per cent. No dependence of the absorption coefficient on the slight variation of temperature (of the order of 1°C) between different experiments was detected.

3. Results

The results of eight satisfactory experiments, weighted with the reciprocal of the variance within each experiment, determine a value for the absorption coefficient with a standard error of 0·18 per cent. To this random error must be added possible systematic errors in the measurement of tube length and of pressure, estimated at 0·21 per cent.

The final result is a value for the absorption coefficient at 3021 Å at 20°C of $3{\cdot}310 \pm 0{\cdot}009$ $(\text{cm STP})^{-1}$.

Ny and Choong's result for this line was 3·72 : Vigroux gives a value for this wavelength (measured with a continuous source) of 3·29. In neither case is an estimate of error given. The present determination is in excellent agreement with Vigroux's value at the same wavelength.

It is hoped to publish later a more detailed account of this work which is being extended to other wavelengths.

Acknowledgments

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Ozone information from sky observations

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(Manuscript received 1 May 1957)

Summary

The information of the vertical distribution of atmospheric ozone obtainable from measurements of scattered sunlight received from various directions in the sky is considered. It is shown that observations in a particular direction in the sky could replace, but would not supplement the customary zenith sky observations, except in high latitudes at certain seasons of the year. Observations in various directions in the sky would be equivalent to zenith sky observations made throughout the day.

1. Introduction

In a recent publication (Walton 1957, to be referred to as I), consideration was given to the problem of determining the vertical distribution of atmospheric ozone from measurements of the clear zenith skylight made with the Dobson spectrophotometer. This instrument could be

arranged to receive scattered light from a direction other than the vertical, or from various directions in the sky in turn. In the present note we examine whether such measurements could usefully replace or supplement zenith sky measurements. We limit the problem to a plane-stratified atmosphere, neglect the effects of secondary and higher orders of scattering, and adopt the notation of I. Additional notation is explained as required.

2. Observations in a particular direction

Considering two wavelengths λ and λ' in a beam of primary scattered light received in any direction making an angle θ with the vertical direction through the observer, when the solar zenith distance is z, the intensity-ratio may be written as follows:

$$\frac{I(\theta,z)}{I'(\theta,z)} = \frac{s\, I_\infty\, e^{-(a+s)\sec\theta} \int_0^1 e^{-(a\,m_0 + s\,m)(\sec z - \sec\theta)}\, dm}{s' I'_\infty\, e^{-(a'+s')\sec\theta} \int_0^1 e^{-(a'm_0 + s'm)(\sec z - \sec\theta)}\, dm}, \quad . \quad . \quad (1)$$

which reduces to Eq. (6) of I when $\theta = 0$ (zenith observations).

Since the Dobson spectrophotometer only measures the log intensity-ratio + const., we consider the ratio of two quantities of the form (1). Thus referring the intensity-ratio to that given by $z = \theta$ we have

$$\frac{I(\theta,z)}{I'(\theta,z)} \Big/ \frac{I(\theta,\theta)}{I'(\theta,\theta)} = \frac{J(\sec z - \sec\theta)}{J'(\sec z - \sec\theta)} = R(\sec z - \sec\theta) \text{ say}, \quad . \quad . \quad (2)$$

where J is the integral defined in Eq. (1) of I, the integrand of which contains the function $m_0\,(m)$ representing the vertical distribution of ozone. The problem is to obtain information of $m_0\,(m)$ from observed values of $\ln R$.

Since (2) is of the same form as Eq. (9) of I, observed values of $\ln R$ plotted against $(\sec z - \sec\theta)$ would reproduce the $(\ln R, \sec z - 1)$ curve, or *umkehr* curve. Hence observations in a direction 'θ' could replace, but would not supplement, zenith observations. Moreover, the intensity of the light received at the instrument is less than that for zenith sky observations (by the factor $e^{-(a+s)(\sec\theta - 1)}$ in the case of wavelength λ), so that the latter are generally to be preferred. But in high latitudes at times of the year when the altitude of the sun does not vary greatly, observations in a direction other than the vertical may be useful. For instance, in latitude 80° at the vernal equinox the zenith distance of the visible sun varies from about 86° to 90°, so that in the case of zenith observations the lowest possible value of $(\sec z - 1)$ would be about 13. The results of I, and the work of Götz, Meetham and Dobson (1934), suggest that the useful part of the *umkehr* curve begins where $\sec z - 1 = 5$, approximately. This part of the curve could be obtained in effect by making observations in a direction given by $\theta = 84°$, for then $(\sec z - \sec\theta)$ would vary from about 5 upwards.

3. Observations in various directions

θ and z have the same meaning as in 2, but we now suppose that θ varies while the sun remains in the same position (z constant). For the present discussion it is convenient to represent the vertical distribution of ozone by function $m^*_0\,(m^*)$ where

$$m^*_0 = 1 - m_0, \text{ and } m^* = 1 - m, \quad . \quad . \quad . \quad . \quad (3)$$

so that (1) becomes

$$\frac{I(\theta,z)}{I'(\theta,z)} = \frac{s\, I_\infty\, e^{-(a+s)\sec z} \int_0^1 e^{-(a\,m^*_0 + s\,m^*)(\sec\theta - \sec z)}\, dm^*}{s' I'_\infty\, e^{-(a'+s')\sec z} \int_0^1 e^{-(a'm^*_0 + s'm^*)(\sec\theta - \sec z)}\, dm^*}. \quad . \quad . \quad (4)$$

If we put

$$J^*(\sec\theta - \sec z) = \int_0^1 e^{-(a\,m_0^* + s\,m^*)(\sec\theta - \sec z)}\, dm^*, \quad . \quad . \quad . \quad (5)$$

and refer the intensity-ratio (4) to that given by $\theta = z$, we have

$$\frac{I(\theta,z)}{I'(\theta,z)} \Big/ \frac{I(z,z)}{I'(z,z)} = \frac{J^*(\sec\theta - \sec z)}{J^{*\prime}(\sec\theta - \sec z)} = R^*(\sec\theta - \sec z) \text{ say}, \quad . \quad . \quad (6)$$

where the primed quantities contain a' and s' in place of a and s.

The argument of 2 may now be applied. Thus observed values of $\ln R^*$ plotted against $(\sec\theta - \sec z)$ would reproduce the $(\ln R^*, \sec\theta - 1)$ curve, which would result if the sun were overhead. For simplicity we shall only further consider this latter case.

When $z = 0$, (6) can be rewritten as

$$J^*(\Theta) / J^{*\prime}(\Theta) = R^*(\Theta), \quad . \quad . \quad . \quad . \quad . \quad (7)$$

where $\Theta = \sec\theta - 1$. From the analogy between (5) and Eq. (1) of I, and between (7) and Eq. (9) of I, it follows that $\ln R^*$ can be expressed as a series of the form

$$\ln R^* = \sum_{n=1}^{\infty} (q^*_n - q^*_n{}') \Theta^n, \quad . \quad . \quad . \quad . \quad (8)$$

where q^*_n and $q^*_n{}'$ are expressions of the type given in Eq. (17) of I, but containing $N^*_{1,0}$, $N^*_{1,1}$ etc., in place of $N_{1,0}$, $N_{1,1}$, etc., where corresponding to Eq. (13) of I, we have

$$N^*_{n-r,r} = (r+1) \int_0^1 m^*_0{}^{n-r}\, m^{*r}\, dm^* \quad . \quad . \quad . \quad (9)$$

By an argument similar to that given in I it follows that from observed values of $\ln R^*$, when Θ is small, we could determine $N^*_{1,0}$, and, by the use of a third wavelength, the quantities $(N^*_{2,0} - N^*_{1,0}{}^2)$ and $(N^*_{1,1} - N^*_{1,0})$. But we find that $N^*_{1,0} = 1 - N_{1,0}$, $N^*_{2,0} - N^*_{1,0}{}^2 = N_{2,0} - N_{1,0}{}^2$, and $N^*_{1,1} - N^*_{1,0} = N_{1,1} - N_{1,0}$. Consequently the information given by observations in various directions in the sky near the sun is equivalent to that given by the low-Z part of the *umkehr* curve discussed in I, and does not supplement it.

It may also be remarked that just as the high-Z part of the *umkehr* curve is the only useful part in practice, the high-Θ part of the $(\ln R^*, \Theta)$ curve may be expected to be the only part of value. To obtain this part of the curve would require observations of the sky near the horizon when the sun is near the zenith and such conditions would not obtain in high latitudes at any time of the year. Such observations might be useful in low latitudes, and would have the advantage over zenith observations that they could be made in a short period of time, thus avoiding the assumption of the constancy of the total ozone amount throughout the day. It should be emphasized that the effects of neglecting secondary scattering in the analysis of such observations are not known. The writer hopes to evaluate these effects in due course.

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551.578.13 : 551.501.81

Variation with height of rainfall below the melting level

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(Manuscript received 15 April 1957)

Summary

3-cm radar signal intensity measurements, made at a range of heights up to 6,000 ft on three days with widespread moderate rain, suggest that there was considerable uniformity of drop-size distribution and of rainfall rate with height once melting of snowflakes was complete and terminal velocity was achieved, and that in these warm-front type situations no significant growth of precipitation was taking place below the melting layer.

1. Introduction

Wexler and Austin (1954) and Boucher, Putnins, Webber and Wexler (1955) have made detailed studies of the variations of radar-signal intensity in the melting layer and in the snow above, but there are few reports available of the measurement of variations of intensity below the melting layer. Yet these are of particular importance. Signal intensities are so sensitive to drop-size distribution that it is very desirable that we should know whether the radar signals with which the radar meteorologist usually has to deal, from precipitation well above the ground, are really representative of rainfall at the ground.

Rigby, Marshall and Hitschfeld (1954) and Mason and Ramanadham (1954) have made theoretical studies of accretion and collision processes below the melting layer which suggest that after a fall of only 1 km in layer-cloud conditions the smaller drops should be significantly depleted and the larger ones increased, but confirmation of these effects is hard to come by. A close knowledge of the complete drop-size distribution is desirable, but reliable measurement of drop-size distribution from aircraft has not yet been achieved, while balloon-borne measurement is difficult and has given unexplained results. The best available alternative is from quantitative radar measurements, and this paper summarizes the results obtained on three occasions of continuous moderate rain, in conditions very close to those assumed by Mason and Ramanadham for their theoretical computations.

2. Nature of radar measurements

The radar-intensity scans were made at the Meteorological Office Radar Research Station at East Hill, 30 mi north-west of London, on the amplitude display of a 3-cm AN/TPS-10 range-height radar. Working at a fixed slant range of 3 mi, steps of 2° in elevation corresponded to steps of 500 ft in height. This technique enabled us to scan down to 500 ft above the ground, and had definite advantages over the use of the same equipment looking vertically, for then the receiver is blocked below 2,000 ft by receiver paralysis. Reliable measurements could not be made below 500 ft because of obstructions such as distant trees, on even the clearest bearings. The calibration of the radar was checked before and after each series of measurements.

At range 3 mi the sampled volume is 200 ft deep in the vertical, and roughly 500 ft square in cross-section. The use of a constant slant range meant that a measurement at the maximum elevation of sweep (22°) was on a sample displaced 1,200 ft horizontally from one made at 2° elevation, but this was regarded as acceptable in continuous rain scans. Any error resulting was probably less than that caused by the wind drift of precipitation through the column in the 3 to 5 min. taken over each scan, for which no allowance was made. The scans were made downward, with a mean 'scanning velocity' of 8 m/sec, which is roughly the terminal velocity of fall of a 3-mm diameter drop.

The attenuation of the 3-cm beam by intervening rain over a 6 mi path-length is less than 1 db for the rain intensities involved, and would not in any case significantly affect relative intensities when scanning at a constant slant range.

Averaging a large number of scans should smooth out irregular variations due to heavier or lighter rain drifting into the beam, but should not remove a regular increase in intensity due to accretion and aggregation during fall.

Results obtained on three days of widespread continuous moderate rain, in warm-front or closely similar conditions, are used in this paper. On each occasion there is evidence from our 10-cm radar records that the rain was reasonably uniform over an area of several hundred square miles. Comparison has also been made with recording-raingauge records at Dunstable, 4 mi from the scanning site, and with records from the nearest hourly weather reporting station at Cranfield 12 mi to the north-west.

3. Typical records

Although only the average of a large number of scans can be expected to reveal a small increase or decrease of intensity with height the individual scans themselves are of much interest, and a typical sample of the material, in this case taken on 20 Oct 1955, is presented.

The synoptic situation can be described briefly. At 1200 GMT on 20 October a deep depression which had crossed Ireland and southern Scotland on 19 October was moving up the Norwegian coast and filling up (pressure at centre 984 mb). A northerly airstream had spread to most parts of the British Isles, but a wave had developed on the trailing front of the depression, and a period of continuous rain was experienced at East Hill during its passage.

Measurements were made in two periods, 0957 to 1023 hr and 1159 to 1227 hr, but only the series of eight scans in the latter period is reproduced in detail (Fig. 1). Signal-intensity measurements were made in steps of 1° through the bright band, which shows up clearly in each scan at about 5,000 ft, and in steps of 2° in the rain beneath. The signal intensity is plotted in decibels above an arbitrary but calibrated level, each successive scan being displaced 5 db to the right to avoid overlapping. Usually only a few seconds elapsed between the completion of one scan and the beginning of the next, each scan taking about 4 min.

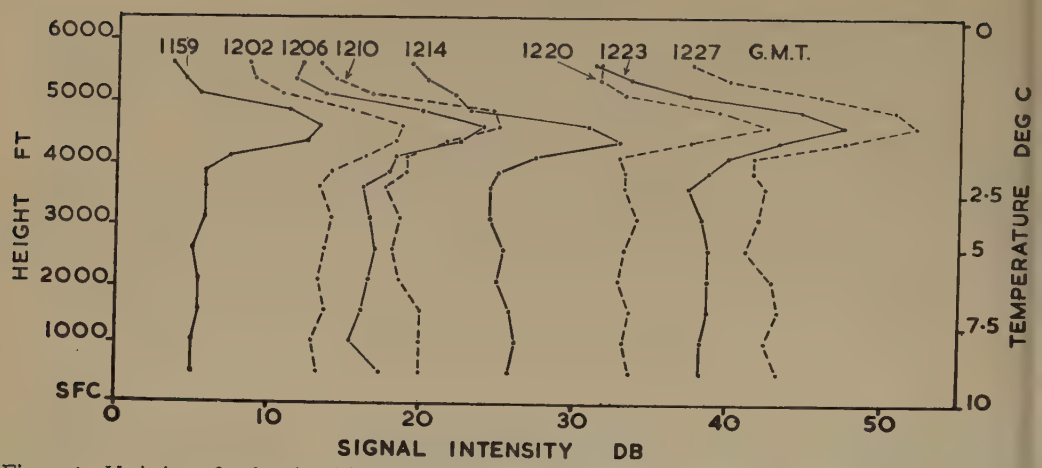

Figure 1. Variation of radar-signal intensity with height throughout a series of 8 scans on 20 Oct. 1955. The temperature scale is based on the nearest radiosonde ascent.

The bright band at the melting layer is seen to be quite uniform in intensity and height, and to be only a little over 1,000 ft in thickness. The intensities below the melting layer show considerable uniformity. Usually a setting would repeat within 1 db on any one scan, and the standard deviation of signal intensity below the bright band over the whole period is $\pm$ 1·7 db.

The temperature scale which has been added at the right-hand edge of Fig. 1 is based on the nearest available radiosonde ascent, made at Hemsby 100 mi to the north-east at 14 hr. The bright-band maximum falls on average 1,500 ft below the Hemsby 0°C isotherm, a rather high value from our experience, probably due to prolonged cooling by melting snow. This is supported by the temperature structure of the Hemsby sounding, which is close to the 0°C isothermal between 4,000 and 6,000 ft, and in which the layer from 0° to 2·5°C is almost 3,000 ft deep, but that from 2·5° to 5°C is barely 1,000 ft deep.

The recording raingauge at Dunstable showed a steady rate of fall of 1·2 mm/hr throughout this period of radar intensity measurements.

4. Discussion

Only 5 of the 32 scans showed a significant increase of signal intensity towards the ground, averaging 2·6 db/km, while three others showed an almost equal decrease, averaging 2·0 db/km. The remaining 24 scans showed no obvious trend towards increase or decrease with height, and it seems most likely that the variation in the 8 cases was due to heavier or lighter rain drifting into the beam in the course of the scan.

Mean values of signal intensity for the full 32 scans have been plotted (Fig. 2, full curve), and a scale based on the mean rate of rainfall at Dunstable (2·7 mm/hr) has been added. (On this logarithmic plot the standard radar equation and the assumption of Best's (1950) drop-size distribution relationship actually determine the scale, the rainfall rate locates it). It cannot be regarded as strictly absolute, but is a useful guide. There is seen to be remarkably little change of intensity with height below the melting level.

Hourly weather reports from Cranfield suggest that the cloud conditions were rather similar in each steady rain period, with variable amounts of thin fractostratus at 500 to 1,000 ft and a main overcast cloud layer with its base at 1,500 to 2,200 ft (one report of 4,000 ft). The relative humidity was between 93 and 99 per cent in each rain period, and it is assumed that evaporation of drops below cloudbase would have been negligible. This is supported by upper-air soundings which were close to saturation to great heights.

Mason and Ramanadham (1954) in their step-by-step computation of accretion and aggregation assumed a steady rate of rainfall of 3 mm/hr, a 1 km fall through cloud of cloud-water content 0·2 g/m^3, and a cloud base 1 km above the ground. These conditions are almost identical with ours, and 0·2 g/m^3 seems a reasonable and modest value for cloud-water content. They computed that 70 per cent of the increase of echo intensity in cloud (1·9 db/km) should be contributed by accretion of cloud droplets, 30 per cent (0·8 db/km) by aggregation of raindrops. Their growth curve, applied at the point A just below the melting layer is the broken line in Fig. 2, with aggregation alone occurring below the assumed cloud base at 2,000 ft.

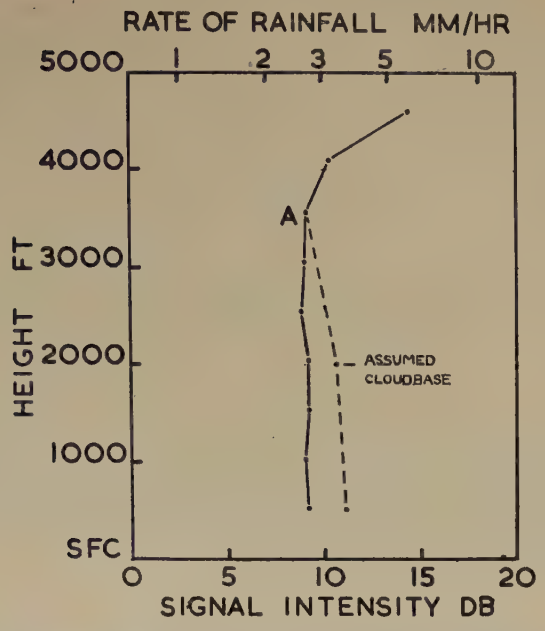

Figure 2. Average signal intensity from 32 scans. The broken line is a computed growth curve. The scale of rate of rainfall does not apply to the curve above the point A.

There is seen to be a clear separation between the computed and measured growth curves. This suggests that the cloud-water content below the melting level was much lower than 0·2 g/m^3. It may well be that the initial rain at a warm front washes out most of the cloud droplets below the melting layer, and that unless there is some cloud renewal mechanism at these low levels (e.g., by topographic or frontal uplift) the cloud-water content thereafter remains low. One would then be forced to assume that the cloud reported at 2,000 ft was a second layer of turbulence cloud of negligible water content, and that the solid cloud deck was actually at about 4,000 ft. This may prove to be typical of warm-front precipitation in those parts of the British Isles in which topographic effects are unimportant, though difficult to observe visually from the ground. It could probably only be reliably confirmed from aircraft.

An increase of 0·8 db/km due to aggregation alone is necessarily more difficult to observe, but is not in fact apparent in Fig. 2. Perhaps we had an unusually narrow spread of drop sizes which resulted in reduced aggregation.

We conclude from the relative constancy of the measured signal intensity with height that there was considerable uniformity in drop-size distribution once melting was complete and terminal velocities had been achieved, and that all significant growth of precipitation was occurring at and above the melting level. It seems that in these conditions the radar measurements at 2,000 ft are an accurate measure of the rainfall at the ground.

Acknowledgment

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The sampling of rain and cloud drops using hygrophotographic plates covered with various kinds of oil

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Summary

The preparation of a hygrophotographic plate is described, gives excellent results in the sampling of cloud droplets, when the latter have diameters between 20 and 30 μ. To avoid the shattering of larger drops (100 to 200 μ) the plate should be smeared with a layer of castor oil. The droplets of water, owing to their density being greater than that of the oil, sink to the bottom of the oily film, are immediately absorbed by the gelatine and thus are recorded.

1. Introduction

In the same way as photography is the art of obtaining images by exposing a sensitive surface to light, hygrophotography is the name given to the process by which an image can be obtained on a sensitive surface by the combined action of light and humidity. The sensitive surface of a hygrophotographic plate has the property of changing its colour when exposed first to light and then to humid atmospheric conditions. It can be prepared from an ordinary photographic plate by impregnating the emulsion, in a manner described below, with a double salt of mercury and silver iodide. A plate, coated in this way, is normally yellow but when exposed to direct or even diffused light of sufficient intensity it changes its colour and becomes blackish-violet. If the exposed plate be then kept in a dark place for a length of time which, depending on the intensity of the blackish-violet colour, may vary between two and several days, it will gradually return to its original yellow colour. This is due to the action of atmospheric water vapour and in fact, if the plate be immersed directly in water, the yellow colouration reappears instantaneously.

2. Preparation of the plates

A hygrophotographic plate is prepared in the following way; an ordinary unexposed photographic plate (coated with a silver bromide emulsion) is immersed in developer solution in a darkroom, then exposed to daylight and finally fully developed. In this way the reduction and blackening are carried out in an extremely uniform manner. The developed plate is then fixed in a solution of hypo to remove traces of unreduced bromide, washed carefully and plunged into a 3 per cent solution of mercuric chloride which bleaches it completely. After again washing the plate to remove excess mercuric chloride it is immersed in a 0.5 per cent solution of potassium iodide. This iodide concentration must be adhered to rigorously because a stronger solution leads to complete blackening of the plate. In the iodide solution the colour of the plate gradually changes to yellow. When a uniform yellow colour has been obtained and the plate is considered, after inspection, to be free from traces of the white chloride, it is taken out of the bath, quickly rinsed in water and allowed to dry (Sivadjian 1953b).

The same process may be used to obtain a hygrophotographic film.

3. Applications of the method in cloud physics studies

(i) Accounts of the use of the hygrophotographic method for the sampling of water droplets have been given by Sivadjian (1952 and 1953a). In the latter reference, an account is given of the application of the method to obtaining life-size images of water droplets, having diameters as small as 20 to 30 μ, on spiders' webs. A typical result is illustrated in Fig. 1.

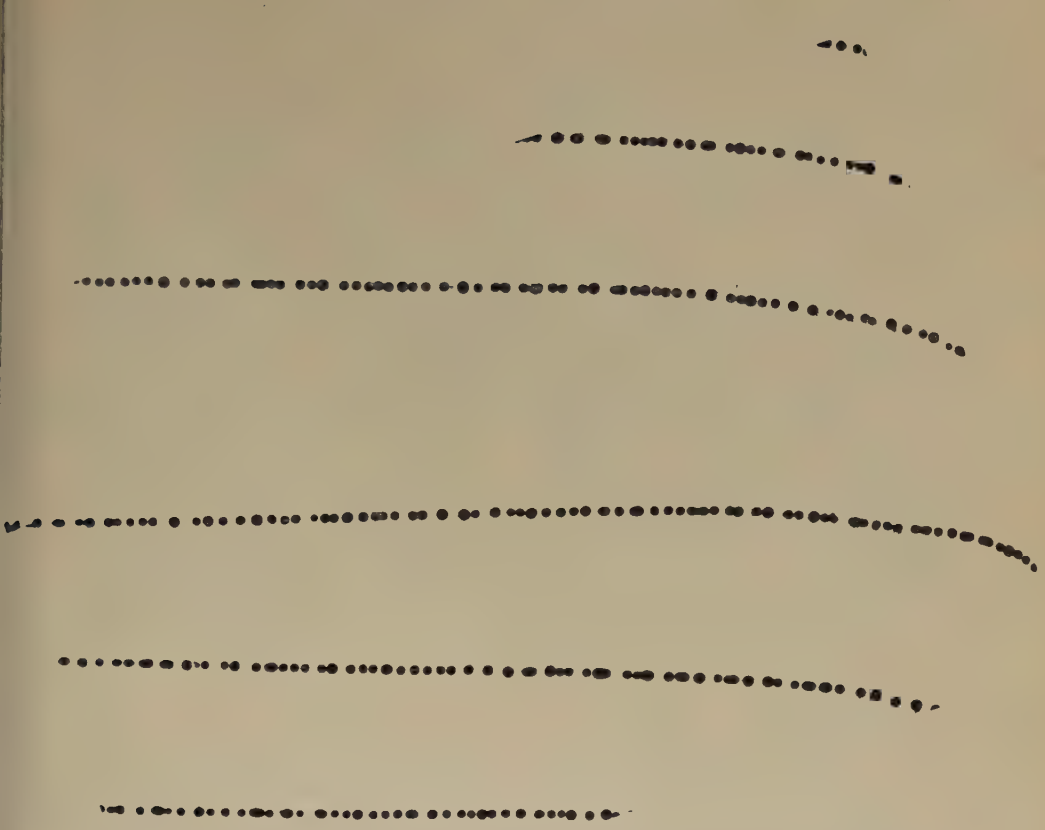

Figure 1. Impressions of droplets on spiders' webs using the hygrophotographic method.

(ii) The method has also been used to sample droplets in natural clouds. This was done from the Hastings aircraft of the Meteorological Research Flight, Farnborough. Individual droplet samples were obtained by mounting a hygrophotographic plate in an impactor, pushing the impactor through a hole in the aircraft's skin and exposing the plate to the cloud droplets by means of a shutter arrangement. At an airspeed of about 180 kt and with an exposure time of 0·01 sec excellent results were obtained for droplets having diameters between 20 and 30 μ. Droplets having diameters between 100 and 200 μ, however, splashed on impact owing to the high speed at which the aircraft was flown.

4. The use of oil-covered plates

A technique to overcome this difficulty is to sample on slides coated with oil. The oil acts as a cushion for the droplets, preventing the fracture of the larger ones and, further, allowing the droplets to retain their natural spherical shape. The use of oil on glass strips for sampling rain or cloud drops was proposed by Hausser and Strobl (1924) following their experiments in which glass strips, coated with a thin layer of castor oil, were exposed to fog droplets. Houghton and Radford (1938) used white vaseline which had the advantage of not slipping along the glass plate.

The use of oil-smeared glass strips is open to criticism however. Boucher (1950) has drawn attention to the rapid evaporation of droplets having diameters less than 5 μ from a glass strip coated with vaseline. He has also referred to the coalescing, due to surface-tension effects, of small droplets on an oil-covered surface.

It was thought that an improved technique might result from a combination of the hygrophotographic plate and oiled-slide methods and an investigation along these lines was undertaken using various kinds of oil.

A hygrophotographic plate, prepared and darkened as previously described, is coated on the darkened sensitive side with a thin layer of oil using a flat brush. It is necessary to wait a little while before exposing the slide to the rain or aerosol droplets, to allow the oil surface to become smooth. Then, having exposed the plate to the droplets, it is again necessary to wait a while, from a few seconds to a few minutes, to allow the droplets to be recorded in the gelatine on the hygrophotographic plate. The oil is then wiped off using a wad of cotton-wool or rag and, if required, final traces of oil can be removed by washing the plate in chloroform. However, for obtaining a satisfactory contact print on bromide paper or a glass plate, cleaning with cotton-wool is sufficient.

It must be stressed that the plates must be completely darkened before applying the oil. In winter or in cloudy weather this is best achieved by exposing them near a window in a well-lit room a long time in advance.

Results obtained using two kinds of oil on different sections of one plate, 9 cm × 12 cm, are illustrated in Fig. 2 (Plate VI). The left-hand section of the plate was coated with ground-nut oil (Fig. 2 (a)), the central strip remained free from oil (Fig. 2 (b)) and the right-hand section was coated with castor oil (Fig. 2 (c)). Another record using castor oil is illustrated in Fig. 3 (Plate VI). The results shown in Fig. 2 were obtained by exposing the plate to rain in a moderate wind on 14 April 1956.

It has been observed that when raindrops, having diameters greater than a certain value, are sampled upon :

(a) a surface covered with olive oil or ground-nut oil, they break up into a large number of smaller drops (Fig. 2 (a)).

(b) a surface free from oil, they are crushed and spread out (Fig. 2 (b)).

(c) a surface smeared with castor or paraffin oil, they retain their own shapes without any deformations (Fig. 2 (c)).

In this last case, the droplets adhere to the oil yet penetrate to the gelatine to be absorbed and are thus recorded. However, the records of the drops obtained using paraffin oil have dimensions distinctly larger than the natural sizes of the drops, whereas using castor oil the exact sizes are preserved with remarkably clear outlines. Using paraffin oil the outlines of the rain-drop records grow and become less clearly defined.

Droplets having microscopic dimensions sink to the bottom of the oily film and, being immediately absorbed by the gelatine, escape risks of coalescence and evaporation and can thus be faithfully recorded.

Acknowledgment

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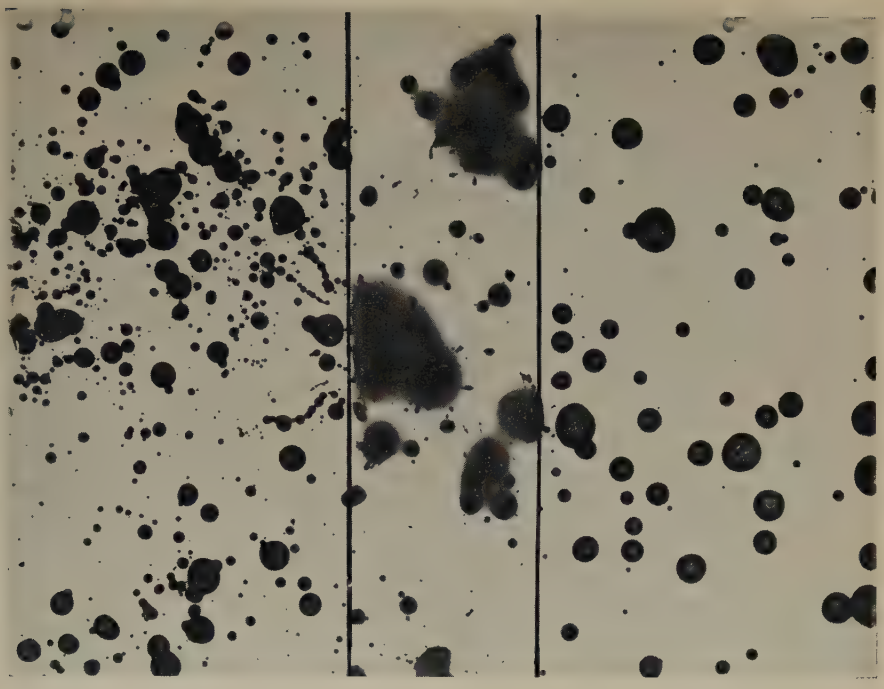

(a) (b) (c)

Figure 2. Impressions of drops obtained on a hygrophotographic plate covered with (a) Ground-nut oil, (b) no oil, (c) castor oil.

Figure 3. Another example of impressions of drops obtained using a coating of castor oil on a hygrophotographic plate.

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On a distribution of weather cycles by length

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Summary

In the paper a weather cycle is defined as the combination of a wet (dry) spell with the successive dry (wet) spell. On the assumption that the lengths of wet and dry spells are independent and obey geometric distributions, the distribution function of the lengths of weather cycles is deduced and tested against observational data for Tel Aviv. The peak frequency of weather-cycle lengths at Tel Aviv is found to be at about 4 days.

1. Introduction

The distribution of the lengths of wet and dry spells* at Tel Aviv, Israel, was considered in a recent paper (Neumann 1955). The results indicate that the length distribution of each type of spell conforms with a ' geometric ' distribution to a good degree of approximation.

The purpose of the present paper is to investigate the distribution of the lengths of weather cycles at Tel Aviv. We define a weather cycle as the combination of a wet (dry) spell with the dry (wet) spell immediately preceding or following it. It is shown elsewhere (Gabriel and Neumann 1956) that the observed length distribution of both spells and weather cycles may be generated by considering a simple stochastic process in which the occurrence or non-occurrence of precipitation on any one day depends only on the occurrence or non-occurrence of precipitation on the immediately preceding day. In this paper, however, we set out from a half-way point; we assume that both wet and dry spells at Tel Aviv ' obey ' geometric distributions.

2. The geometric distribution

Let X be a positive, integral-valued, random variable. We say that X obeys a geometric distribution (see, e.g., Feller 1950, p. 217) if

$$Pr\{X = k\} = q' p'^{k-1}, \qquad (1)$$

$k = 1, 2, \ldots, q'$ and p' being positive numbers such that $p' + q' = 1$.

Applied to wet and dry spells, (1) may be given the following figurative interpretation. Let p' be the probability of a wet day to be followed by a wet day. Then, under the assumption stated above, the probability of a wet day to be followed by $k - 1$ wet days is p'^{k-1}. The complement of p', $1 - p' = q'$, evidently is the probability of a wet day to be followed by a dry day. Since every wet spell has a last day, $(1 - p')$, also is the probability of a wet spell to terminate. Hence, the probability of a wet spell to last *exactly* k days is

$$(1 - p') p'^{k-1}, \qquad (2)$$

which is identical with (1).

In the foregoing paragraph, the term ' wet ' can, of course, be replaced by the term ' dry.' However, the probability of a wet day to follow a wet day need not be the same as the probability of a dry day to follow a dry day.

Let the random variable X represent the length of wet spells and the random variable Y the length of dry spells. We shall denote by p' and p the corresponding associated probabilities. Then,

$$\text{(a)} \quad Pr\{X = k\} = (1 - p') p'^{k-1} = a_k \qquad (3)$$

$$\text{(b)} \quad Pr\{Y = m\} = (1 - p) p^{m-1} = b_m$$

$k, m = 1, 2, \ldots$.

* A wet (dry) spell is defined as a continuous run of k days, $k = 1, 2, \ldots$, each (none) of which is credited with a measurable amount of precipitation.

3. The length distribution of weather cycles

We shall now assume that the variables X and Y are independent, i.e., we assume the lengths of wet (dry) spells to be uncorrelated with the lengths of dry (wet) spells preceding or following them. Actually, this is not a new assumption for, as stated briefly in the introductory section, the geometric distribution of spells may be generated from the simple stochastic model according to which the wet or dry character of any day depends on the character of the preceding day only. According to this model, then, the lengths of wet and dry spells would be independent.

Let Z be the variable representing the lengths of weather cycles, $Z = X + Y$. Since both X and Y are random variables, so also is their sum Z. Moreover, $X + Y = Y + X$, that is, it should be immaterial if weather cycles are considered to begin with a wet or dry spell: their length distribution should be the same.

The event $Z = n$, $n = 2, 3 \ldots$, is the combination of the events $(X = 1, Y = n - 1), \ldots, (X = n - 1, Y = 1)$. Having assumed that X and Y are mutually independent and since the events $(X = 1, Y = n - 1), \ldots, (X = n - 1, Y = 1)$ are exclusive, the probability of cycles n days long is obtained as the sum of all the products of probabilities of wet spells of k days and dry spells of m days such that $k + m = n$. From (3).

$$Pr\{Z = n\} = a_1 b_{n-1} + \ldots + a_{n-1} b_1,$$

$$= c_n, \text{ say.} \quad . \quad . \quad . \quad . \quad . \quad (4)$$

From (3) and (4)

$$c_n = (1 - p)(1 - p')(p^{n-2} + p^{n-3} p' + \ldots + pp'^{n-3} + p'^{n-2}). \quad . \quad (5)$$

On the right-hand side, the long bracketed term can be summed. Writing $f(n; p, p')$ for c_n we get,

$$f(n; p, p') = (1 - p)(1 - p') \frac{p^{n-1} - p'^{n-1}}{p - p'} \quad . \quad . \quad . \quad (6)$$

(6) is the distribution function of the length of weather cycles. In form it is symmetric in p and p', a result one would expect in view of the commutative property of the sum of the random variables X and Y.

If $p = p'$, a passage to the limit $p' \to p$ reduces (6) to

$$f(n; p) = (n - 1)(1 - p)^2 p^{n-2} \quad . \quad . \quad . \quad . \quad (7)$$

which is a special case of a well-known distribution in probability theory, viz. the so-called Pascal distribution or negative binomial distribution (e.g., Feller 1950, p. 218) arising from a combination (' convolution ') of two identical ($p = p'$) geometric distributions. It is for this reason that we suggest calling (6) ' generalized Pascal ' or ' generalized negative binomial ' distribution, (p and p' arbitrary except that $1 > p, p' > 0$).

The distribution function (6) is, in the most general case, unimodal of the bell-shaped type. To show this, write $n = 2i - 1$, $i = 2, 3, \ldots$ and consider the ratio

$$\frac{f(2i - 1; p, p')}{f(i; p, p')} = p^{i-1} + p'^{i-1} \quad . \quad . \quad . \quad . \quad (8)$$

Since p and p' are probabilities and $1 > p, p' > 0$, the greatest value of the ratio is reached for the lowest admissible value of i, viz. 2 (in this paper, weather cycles may not be shorter than two days) when $p^{i-1} + p'^{i-1} = p + p'$. Now, in the meteorologically relevant case, $p + p' > 1$. For if $p + p' \leqslant 1$, $p' \leqslant 1 - p$, and this would mean that the probability of a wet day to follow a wet day is less than, or at most equal to, the probability of a wet day to follow a dry day and, in turn, such a relationship would imply that a given type of weather tends to change, or at most, shows no sign of persistence, contrary to physical facts. Therefore, throughout this paper, we assume that $p + p' > 1$.

Even though for $i = 2$, the smallest admissible value of i, the right-hand member of (8) is greater than 1, there will be a value of i, $i > 2$, for which the ratio (8) eventually becomes less than 1. The meaning of this is that the function $f(n; p, p')$ increases at first, and then onward from some value of i, it decreases, i.e. the discrete distribution has a maximum. And because $p + p' > 1$, it is clear from (8) that the shortest cycle length at which the mode can occur is at $n = 3$, or *the most frequent length of weather cycles cannot be less than three days.*

Strictly, it is possible that the maximum frequency, or modal frequency, is equally shared between two consecutive cycle lengths. Suppose that the modal length of weather cycles is n' and that, possibly, the modal frequency is equally shared by cycle lengths n' and $n' + 1$, then, from (6).

$$p^{n'-2} - p'^{n'-2} < p^{n'-1} - p'^{n'-1}, \qquad p^{n'-1} - p'^{n'-1} \geqslant p^{n'} - p'^{n'}.$$

It is readily found from these inequalities that the modal length n' is the integer located between the following numbers differing by unity,

$$\frac{\log (1 - p')/(1 - p)}{\log (p/p')} + 1 \leqslant n' < \frac{\log (1 - p')/(1 - p)}{\log (p/p')} + 2. \quad . \quad . \quad (9)$$

or if the two numbers are integers then n' equals the smaller of the two.

We close this section by stating the mean and variance of distribution (6). Let μ', μ, σ'^2, σ^2 be the means and variances of the lengths of wet and dry spells and μ_c and σ_c^2 the mean and variance of the lengths of cycles. The mean and variance of geometric distributions are related to the fundamental probabilities p, or p', by the equations (see Feller 1950, p. 217).

$$\mu' = E(X) = \frac{1}{1 - p'}, \qquad \mu = E(Y) = \frac{1}{1 - p} \quad . \quad . \quad (10a)$$

$$\sigma'^2 = \mathrm{Var}(X) = \frac{p'}{(1 - p')^2}, \quad \sigma^2 = \mathrm{Var}(Y) = \frac{p}{(1 - p)^2} \quad . \quad . \quad (10b)$$

Because (6), as well as (7), is generated by the combination (' convolution ') of two geometric distributions, assumed independent, the mean and variance of (6) and (7) are the sums of means and variances, respectively, of (3), i.e., of the spells. From (10),

$$\mu_c = E(X + Y) = E(Z) = \frac{1}{1 - p'} + \frac{1}{1 - p}$$

$$\sigma_c^2 = \mathrm{Var}(X + Y) = \mathrm{Var}(Z) = \frac{p'}{(1 - p')^2} + \frac{p}{(1 - p)^2}.$$

4. Comparison of results with observed data

In order to check how far (6) represents the observed length distribution of weather cycles, we shall present data for Tel Aviv (Nahmani Street). The data relate to the months December, January and February, i.e., the winter season, during which we find (Gabriel and Neumann 1956) the probabilities p and p' show very little variation. The observations cover the 27 years from 1923-24 to 1949-50.

Of course, spells do not begin, in general, on 1st December, nor will 28 (or 29) February always be the last day of a spell. To avoid ' breaking ' the spells, the following procedure was adopted: (a) wet spells were allocated to the winter season if their first day fell in the period 1 December to 28 February (or 29) even though in the latter case they may have ended early in March; (b) dry spells were counted if the immediately following wet spells were included in the winter season as under (a); (c) cycles were allocated as their wet spells.

The principal data are listed in Table 1, adding values of the probabilities p and p' from (10a), taking for the ' population ' mean spell lengths μ' and μ the corresponding sample means. It is possible to show, by the maximum likelihood principle, that the best estimates of population means are the sample means.

TABLE 1. Wet and dry spells at Tel Aviv in winter (December, January and February), 1923-24–1949-50

No. of years of record	No. of days wet	No. of days dry	No. of spells	Probability (= relative frequency) p' (wet)	Probability (= relative frequency) p (dry)	Goodness of fit (χ^2-test of (3)) probability wet spells	Goodness of fit (χ^2-test of (3)) probability dry spells
27	1037	1400	351	0·662	0·750	0·30-0·40	0·50-0·60

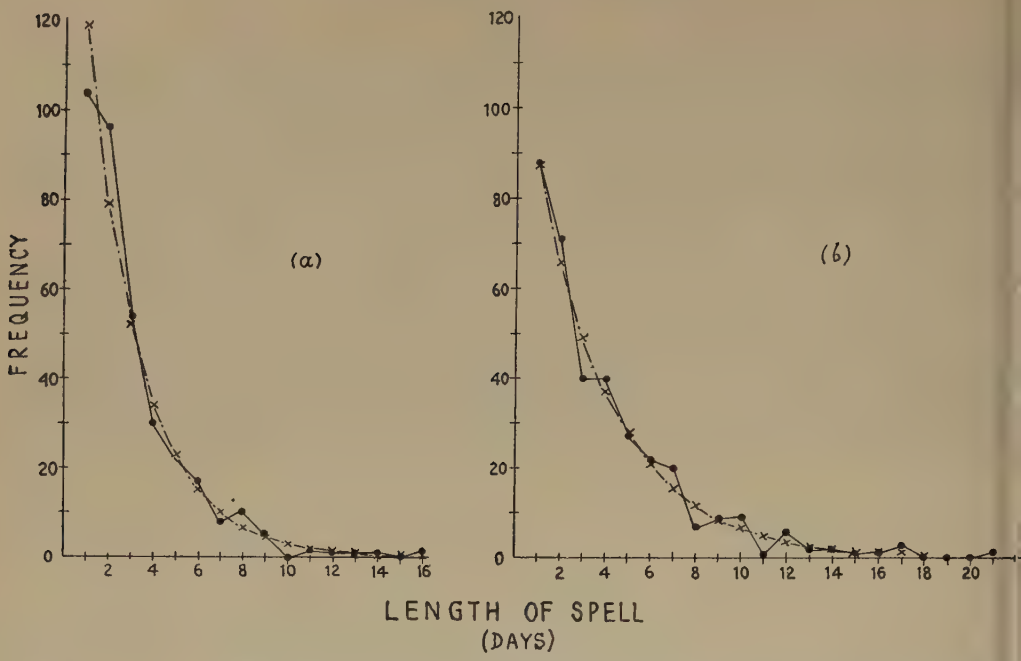

Figure 1. Observed and theoretical distributions of lengths of (a) wet spells, and (b) dry spells, at Tel Avi in winter, (December, January and February), 1923-24–1949-50.

Dots : Observed spell lengths. Crosses : Theoretical spell lengths computed from (3).

Figures 1 (a)-(b) show the observed length distributions of wet and dry spells compared with the theoretical* distributions calculated from (3) using the values of p and p' quoted in Table 1. It is seen that the fit is quite good, a fact corroborated by the χ^2-test, the results of which are stated in Table 1. For wet and dry spells the chances are approximately 35 per cent and 55 per cent, respectively, that differences as large as or larger than those between the theoretical and observed frequencies could have arisen by chance.

Before calculating the theoretical distribution of cycle lengths, it is appropriate to check if the lengths of wet and dry spells are independent. The linear correlation coefficients between the lengths of successive wet and dry spells as well as those between the lengths of successive dry and wet spells in Tel Aviv have been presented elsewhere (Gabriel and Neumann 1956) for each of the months November to April inclusive. Table 2 cites the figures for the months December to February.

Table 2 indicates that in each case the correlation is low. It also tests, by Fisher's z transformation, the significance of the coefficients and it is seen that none of the correlations reaches the 5 per cent significance level. Hence, it is warranted to regard, to an approximation, the lengths of alternate kinds of weather as independent.

TABLE 2. LINEAR CORRELATION OF THE LENGTHS OF SUCCESSIVE SPELLS OF WET AND DRY WEATHER AT TEL AVIV (27 YR) (r = LINEAR CORRELATION COEFFICIENT)

Month	No. of pairs of spells	Wet-dry pairs r_{WD}	Wet-dry pairs Unit normal deviate	Dry-wet pairs r_{DW}	Dry-wet pairs Unit normal deviate
December	117	− 0·090	− 0·96	0·008	0·09
January	133	− 0·042	− 0·48	− 0·148	− 1·61
February	101	0·045	0·45	− 0·030	− 0·30

* Here and below, the term 'theoretical' refers to calculations based on Eqs. (3) and/or (6).

Introducing the values of p and p' (Table 1) into (6), gives the theoretical lengths of weather cycles at Tel Aviv. These are listed in Table 3 and graphed in Fig. 2. The mode of the theoretical distribution is at 4 days, a result one could obtain directly from (9). Further, Table 3 and Fig. 2 show the observed length distribution of cycles beginning with a wet spell (' wet-dry ' cycles) and that of cycles beginning with a dry spell (' dry-wet ' cycles).

TABLE 3. THE LENGTH DISTRIBUTION OF WEATHER CYCLES AT TEL AVIV IN WINTER (DECEMBER, JANUARY AND FEBRUARY), 1923-24—1949-50.

No. of days in cycle	Observed No. of wet-dry cycles	Observed No. of dry-wet cycles	Theoretical No. of cycles (from (6))
2	33	30	30
3	38	31	42
4	33	53	44
5	50	40	42
6	39	39	37
7	26	37	32
8	30	19	26
9	24	27	21
10	16	14	17
11	14	16	14
12	15	11	11
13	9	6	8
14	3	9	6
15	3	6	5
16	8	4	4
17	1	2	3
18	4	2	2
19	2	1	2
20			1
21	2	1	1
22		1	1
23	1		1
24		1	
25 +		1	1
Total	351	351	351

It is a curious fact that the observed distribution of wet-dry cycles differs to some extent from the distribution of dry-wet cycles. For example, the observed mode of dry-wet cycles is at 4 days, in agreement with the theoretical value, whereas the observed mode of wet-dry cycles is at 5 days. Despite the seeming disagreement between wet-dry and dry-wet cycles, it is found that their fit is about equally good. Application of the χ^2-test shows that for both wet-dry and dry-wet cycles, the probability is 40-50 per cent that differences as large as, or larger than, those between the observed and theoretical frequencies, might have arisen by chance.

5. Discussion

In Section 3 we have inferred that the modal length of weather cycles cannot be less than 3 days. In Section 4 the data of Tel Aviv have been examined and it was found that the modal length of the theoretical distribution, as well as that of the observed distribution of the dry-wet cycles, is 4 days. It is, therefore, of some interest to point out that the occurrence of a peak frequency of about 4 days in the length distribution of weather cycles is not restricted to Tel Aviv. Longley (1953), who favours the view that the lengths of wet and dry spells of any given season conform closely with geometric distributions, published data either of p' and p or such other data from which p' and p can be readily computed. Longley's data comprise five stations in Canada (Dawson, Yuk.; Montreal, P.Q.; Saint John, N.B.; Victoria, B.C.; and Winnipeg, Man.), one station in the United States (San Francisco, Calif.), and four stations in the British Isles (Aberdeen, Greenwich, Kew, Valencia). When the appropriate values of p' and p are introduced into (9) above, it is found that for all these stations the peak frequency is for a cycle length of between 3 and 5 days.

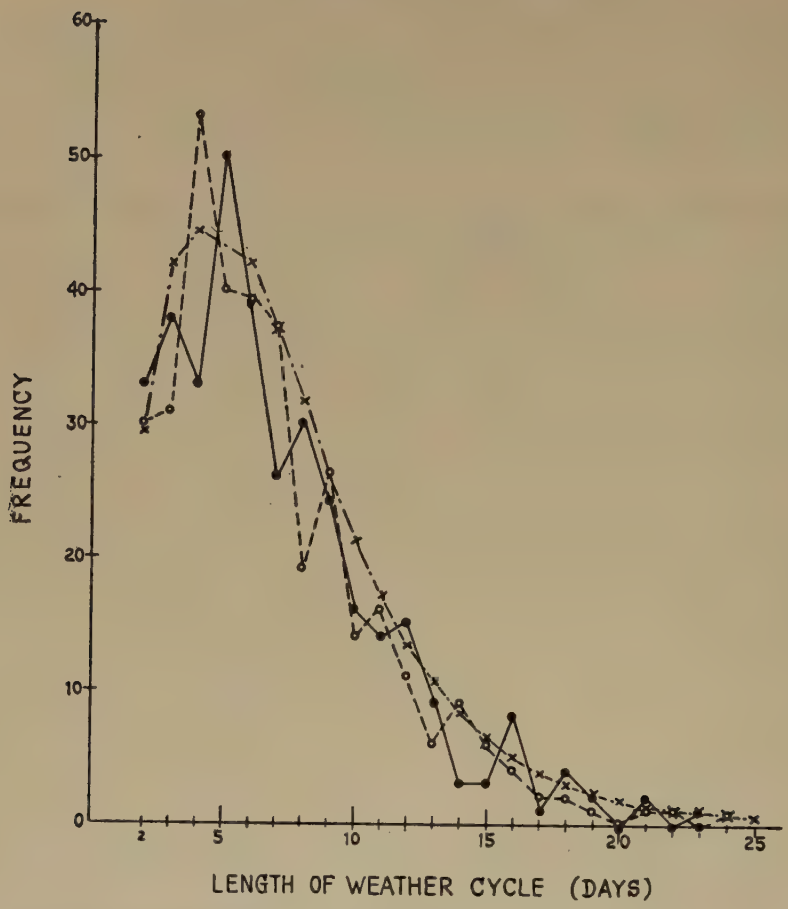

Figure 2. The length distribution of weather cycles at Tel Aviv in winter (December, January and February) 1923-24–1949-50.

Full circles: wet-dry cycles. Open circles: dry-wet cycles. Crosses: theoretical cycle lengths computed from (6)

The above findings are further supported by the recent results of Estoque (1955) who studied the time spectrum of intensity of the turbulent transfer of heat and momentum. An application of his theoretical results to data for the 850-mb level over the south-eastern United States shows that the most important transfers occur by disturbances of a ' period ' of 4 days (see Estoque Fig. 4, p. 183). Rather similar results emerge from an even more recent investigation of the spectrum of air temperature at University Park, Pa., by Griffith, Panofsky and Van der Hoven (1956). These authors find that the spectrum has a large peak at 4 days and suggest that this peak reflects, perhaps, the average time between frontal passages. It is interesting to note that the present paper predicts a similar peak on the basis of simple probability considerations.

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551.557.36 : 551.587

Wind statistics at Singapore

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(Manuscript received 21 January 1957)

SUMMARY

Statistics, computed from radar-wind observations made at Singapore during a period of 5½ years, are presented, and discussed.

1. Data used and statistics

Weekly radar ascents at 02 GMT commenced in July 1950 at Singapore. This frequency increased fairly steadily to about three a week by July 1953, when daily ascents at 03 GMT were commenced. From August 1953 ascents were also made daily at 15 GMT. There was one long break from 4 Dec. 1954 to 16 Jan. 1955. All 02 and 03 GMT ascents made from July 1950 to the end of December 1955 were used in computing the statistics. Since observations of wind at 70,000 ft were relatively infrequent, 15 GMT observations at this height were used in addition to those at 02 and 03 GMT.

For each month (except at 70,000 ft) and season the statistics v, ϕ, θ, σ_x, σ_y, σ, α, and N were computed.

v = speed of vector mean wind in knots.

ϕ = direction of vector mean wind in degrees from north.

θ = angle in degrees between major axis of ellipse representing the wind distribution and the east-to-west direction (Scott 1956a). Its sign is that of the usual trigonometrical convention.

σ_x = standard deviation in knots of components parallel to major axis of the ellipse.

σ_y = standard deviation in knots of components parallel to minor axis of the ellipse.

σ = standard vector deviation.

α = the probability of the distribution being obtained as a sample from a circularly-distributed population (Mauchly, quoted by Brooks and Carruthers 1953). In computing the values of α the number of independent observations was taken to be the maximum number separated by at least one week.

N = the number of observations.

It seemed unlikely, in view of the results found by Scott (1956b) for Singapore, that wind observations separated by two days (Brooks and Carruthers 1953) were independent. Serial correlation tests were therefore made on several sequences. These indicated that two days is probably more than a sufficient interval for independence of S-N components, but is insufficient for W-E components. For these latter the interval had to be increased to one week before an acceptably small serial correlation coefficient was obtained. This interval was therefore taken as the minimum required for independence of a series of W-E components and wind vectors, while two days was considered sufficient for S-N components.

The distribution of components was tested at 10,000, 20,000, 30,000, 40,000, 50,000 and 60,000 ft in the December to February season and at 70,000 ft in the March to May, and September to November seasons. Where the total number of independent observations was relatively small suitable observations made since the end of 1955 were used to augment those made previously.

TABLE 1. WIND STATISTICS AT SINGAPORE (1950 to 1955)

A = December to February; B = March to May; C = June to August; D = September to November.

Season	ϕ	v	θ	σ_x	σ_y	σ	α	N	Height (ft)
A	359	7·3	+ 22	9·5	5·9	11·2	0·008	205	
B	302	3·4	+ 24	8·1	6·2	10·2	0·192	209	5,000
C	232	8·9	− 5	7·4	6·3	9·7	0·673	272	
D	241	6·8	− 20	9·9	6·7	11·9	0·008	308	
A	338	2·9	+ 8	10·1	6·4	12·0	0·002	224	
B	301	2·8	+ 12	8·4	6·3	10·5	0·087	241	10,000
C	255	6·5	− 2	9·5	7·6	12·1	0·147	298	
D	258	4·7	− 9	11·2	6·9	13·1	10^{-4}	324	
A	043	2·2	+ 7	9·9	5·8	11·5	0·001	226	
B	065	1·1	− 4	8·2	6·2	10·3	0·106	242	15,000
C	180	1·3	+ 7	10·3	6·8	12·4	0·004	300	
D	269	0·5	− 2	12·3	6·2	13·8	10^{-7}	328	
A	083	4·2	+ 9	10·4	6·5	12·2	0·003	227	
B	082	4·1	+ 7	8·0	5·7	9·8	0·043	241	20,000
C	092	8·7	− 9	8·8	5·9	10·6	0·004	301	
D	091	5·4	+ 1	10·2	5·4	11·5	10^{-6}	329	
A	080	8·0	+ 4	10·7	7·1	12·9	0·012	228	
B	085	7·3	− 3	7·8	5·9	9·7	0·066	240	25,000
C	092	16·6	+ 9	8·1	5·9	10·0	0·035	299	
D	091	10·3	+ 2	8·4	5·5	10·0	0·001	323	
A	097	9·6	− 11	10·0	7·8	12·7	0·237	229	
B	092	10·0	− 16	7·5	6·5	9·9	0·461	242	30,000
C	085	23·0	− 6	9·0	7·0	11·4	0·085	294	
D	089	15·1	+ 8	9·6	6·3	11·5	0·002	328	
A	112	22·4	− 13	13·1	8·8	15·8	0·025	226	
B	095	17·0	− 6	12·7	10·4	16·4	0·031	244	40,000
C	074	38·6	+ 56	12·3	11·5	16·9	0·670	289	
D	081	27·1	+ 31	14·1	10·0	17·3	0·014	322	
A	100	32·1	− 21	26·2	11·4	28·6	10^{-7}	222	
B	074	18·7	+ 3	20·7	10·7	23·4	10^{-5}	241	50,000
C	078	42·2	+ 28	20·9	12·6	24·4	0·005	260	
D	083	38·1	+ 12	21·5	11·2	24·3	10^{-6}	310	
A	272	4·9	+ 7	14·7	9·3	17·4	0·010	192	
B	041	0·7	+ 5	16·6	8·6	18·7	10^{-5}	212	60,000
C	058	1·3	− 4	15·5	8·9	17·9	0·001	168	
D	256	2·3	+ 3	18·7	9·4	21·0	10^{-6}	233	
A	277	5·3	− 2	13·3	9·1	16·2	0·172	109	
B	091	15·2	− 2	23·9	8·6	25·4	10^{-6}	109	70,000
C	084	13·7	− 1	36·8	8·3	37·7	10^{-9}	53	
D	299	4·3	+ 1	31·9	9·4	33·3	10^{-8}	81	

2. DISCUSSION OF RESULTS

The values of χ^2 found from the distribution tests gave no reason to suppose that all sets of W-E and S-N components up to and including 60,000 ft are not normally distributed. Both sets of W-E components at 70,000 ft however produced excessive values of χ^2. These two distributions appeared distinctly bi-modal. α is less than 0·05 for 29 of 40 seasonal distributions. The proportion was considerably lower for the monthly distributions. However, in only 6 of 108 distributions did θ lie outside the range $\pm$ 35°. These 6 all had relatively high values of α. This suggests that increased volume of data will increase the proportion of monthly distributions showing significant non-circularity.

The empirical rule $\sigma\rho$ = constant (ρ is density) up to the tropopause (Brooks and others 1950) applies from 30,000 ft (the height at which winds are nearest to being circularly distributed) to 50,000 ft. Up to 30,000 ft σ varies little.

Except where $\phi + \theta$ is nearly an integral multiple of 90° the distribution parameters show relation between wind speed and direction, e.g., at 5,000 ft in the December to February season the orientation of the distribution ellipse shows a tendency for higher wind speeds to be associated

with directions veered from that of the mean, and for lower wind speeds with directions backed from the mean. Similarly it is seen that above-average rates of meridional flow at 50,000 ft in December to February, and June to August tend to occur in association with above average east to west components.

Acknowledgments

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Correspondence

551.557.35 : 551.509.317

Weather and upper winds at Singapore

By F. A. Sharp

Scott (1956) has suggested there is a relationship between the 50,000-ft wind and the weather at Singapore. There can be no doubt that the weather is determined by the total wind field but a very partial relationship such as this must be examined critically, especially when it has little physical basis.

Rather than use the 50,000-ft wind it would have been preferable to use the maximum wind which usually occurs near this level. It has more claim to be associated with the weather since it gives a direct indication of the shear just below the tropopause and it also indicates the presence of the ' minor easterly jet ' discussed by Koteswaram (1956).

Let an α day be one on which the maximum wind recorded by the 0300 GMT Singapore radar wind was not less than 70 kt, or not less than 50 kt, if for two previous days it had been not less than 70 kt. Table 1 for the period June-September 1955-6 shows the distribution of total duration of rainfall at Changi, Seletar and Tengah, three stations on Singapore Island. Rain before and after noon zone time has been treated separately. (Zone time = GMT + 7 hr).

TABLE 1

Rainfall Total duration for 3 stations Hours	0000-1200 zone time				1200-2400 zone time			
	α days		Other days		α days		Other days	
	No.	per cent	No.	per cent	No.	per cent	No.	per cent
0	40	53·4	69	42·1	44	58·6	82	50·0
> 0 < 1	13	17·3	34	20·8	13	17·3	37	22·6
⩾ 1 < 2	7	9·3	20	12·2	5	6·7	14	8·5
⩾ 2 < 3	3	4·0	12	7·3	5	6·7	8	4·9
⩾ 3 < 6	4	5·3	14	8·5	5	6·7	12	7·3
⩾ 6 < 9	2	2·7	8	4·9	2	2·7	7	4·3
⩾ 9	6	8·0	7	4·2	1	1·3	4	2·4
Total	75	100·0	164	100·0	75	100·0	164	100·0

There is some evidence in Table 1 that α days are more likely than non-α days to be completely dry, both before and after noon. The evidence becomes weaker if 12 hr periods with less than 20 min average rainfall per station are considered and disappears when 40 min is used as the criterion.

The average duration of rainfall per station when rain did fall was 1·5 hr for α mornings, 1·0 hr for non-α mornings, 0·9 hr for α afternoons and 0·8 hr for non-α afternoons.

From this it might be concluded that the α effect was slightly rain-making in the morning and was nil in the afternoon. It thus becomes clear that the conclusions drawn are very sensitive to the criteria used and that a slight change in the latter may lead to a complete reversal of the conclusion.

As Sharp (1957) has pointed out, the bulk of Singapore rain is of less than 3 hr duration; on only 13 mornings and 5 afternoons did the average duration per station exceed this. Rainfall at Singapore is therefore unlikely to be associated with a wind recorded 48 hr previously and the introduction of such a wind into the criteria is more likely to mask than to reveal a relation between wind and weather. In what follows, therefore, maximum wind on the day in question only is used.

Table 2 represents an approach to the problem which avoids the use of arbitrary criteria. It is a frequency distribution of maximum wind speed at 0300 GMT for various total durations of rainfall at the same three stations for the same period.

TABLE 2

Max Wind speed (kt)	0000-1200 zone time No. of occasions with total duration 0	< 3 hr	⩾ 3 hr	1200-2400 zone time No. of occasions with total duration 0	< 3 hr	⩾ 3 hr	Total distribution
23- 32	4 (3)	2 (2)	– (1)	4 (3)	2 (2)	0 (1)	6
33- 42	15 (13)	10 (11)	4 (5)	11 (15)	11 (10)	7 (4)	29
43- 52	23 (26)	26 (22)	9 (10)	23 (30)	27 (20)	8 (8)	58
53- 62	13 (18)	17 (15)	9 (7)	22 (20)	10 (13)	7 (5)	39
63- 72	18 (21)	18 (17)	9 (8)	31 (24)	11 (15)	3 (6)	45
73- 82	20 (16)	9 (13)	7 (6)	19 (19)	15 (12)	2 (5)	36
83- 92	12 (9)	6 (7)	2 (3)	12 (10)	4 (7)	4 (3)	20
93-102	3 (3)	2 (2)	1 (1)	3 (3)	2 (2)	1 (1)	6
103-112	1 (1)	– (1)	– (0)	1 (1)	0 (1)	0 (0)	1
113-122	1	–	–	0	1	0	1
Total	110	90	41	126	83	32	241

The figures in brackets are the number of occasions which would have been found had each column the same distribution as the total distribution. It is reasonably clear without statistical analysis that each column can be taken as a random sample of the total distribution.

Each of the total durations of rain both before and after noon has the same wind distribution associated with it and weather is therefore independent of the maximum wind speed, which by itself is of no practical forecasting value.

Acknowledgment

The author wishes to thank the Director of the Meteorological Office for permission to publish this letter.

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Meteorological Office,
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Singapore.
10 January 1957.

Reply

By J. R. Scott

The most important difference between F. A. Sharp's investigation and mine seems to me to be that between the weather criteria used. In testing a hypothesis concerning fairly widespread low-level divergence I felt it desirable to take as much direct account as possible of rainfall amounts and spatial distribution; factors which are extremely variable over Singapore. Hence my use of 11 gauges and criteria based on amounts of total fall. Sharp, on the other hand, selects duration meaned over 3 stations only. There is, of course, association between duration of and amount of fall in individual periods of rain, but experience and statistical test suggest that it is not close.* (The correlation coefficient between these quantities, for all (117) individual periods of rain occurring in p.m. periods in June to September 1955 and 1956 at the two most widely separated Singapore rain gauges, was 0·45). Our weather criteria therefore appear to differ more than slightly, and I cannot conclude that his could be more representative of Singapore weather than mine.

* It is relevant to note that drainage engineers usually assume that rainfall intensity is inversely proportional to the sum of duration and a constant.

Another difference which cannot with reasonable certainty be dismissed as slight is that between the wind criteria used. In a critical examination of my suggestion it would have been preferable not to use a criterion which is not obviously closely related to the original. Also, had Sharp used, in computing his Table 2, the wind on the previous day, he would have been somewhat closer to the original investigation. However, he rejects the use of a wind earlier than that on the current morning (for reasons which possibly merit more space).

I have now applied my criteria to p.m. periods in June to November 1955 and 1956. There was one minor difference. This was the use of 10 gauges instead of 11 with a consequent minor change in weather criteria (a period with more than 0·75 in. total fall with 8 gauges recording rain was put into the W class). The results are given in Table 1, presented in a way which permits comparison with Sharp's results. Also given are figures for October and November 1953 and 1954 obtained from the original test.

TABLE 1. FREQUENCIES OF WEATHER TYPES IN P.M. PERIODS DURING JUNE TO NOVEMBER AT SINGAPORE ON α DAYS AND ALL DAYS

Day Class	D	N	W	
α days	54	11	5	June to September 1955 and 1956.
All days	156	59	29	
α days	10	2	1	October and November 1955 and 1956.
All days	56	34	32	
α days	15	3	5	October and November 1953 and 1954.
All days	44	34	44	

The contents of Table 1 give no reason to suppose that the results of the first investigation are misleading.

Meteorological Office,
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10 May 1957.

551.577.11

THE COALESCENCE OF WATER DROPS

By I. C. BROWNE and J. F. P. HUDSON

Cloud physicists often assume that every collision between cloud droplets leads to coalescence. Several experimenters have tested this assumption, but their results are apparently inconsistent. A recent review (Browne, Palmer and Wormell 1954) tried to reconcile the results, and its conclusions are similar to those reached by Swinbank (1947) in a paper, giving experimental and theoretical results, which was unfortunately omitted from the review. The problem is stated concisely by Swinbank: ' Two small drops will assume the shape of minimum surface energy, a sphere. If they coalesce, the joint drop will again be a sphere in which the surface energy has a new and lesser minimum. During amalgamation, the drops will be distorted and .... the distortion may cause an increase in surface energy. Such an increase of surface energy .... would constitute a barrier to coalescence.' From dimensional arguments, Swinbank and Browne, Palmer and Wormell conclude that large drops will coalesce more readily than small drops.

The purpose of this letter is to suggest that it is at least possible for two drops to coalesce in such a way that the total surface area decreases steadily during coalescence, so that there is no energy barrier to the process once it has started. Consider two drops of equal initial radius a, coalescing in such a way that at any instant the liquid in each drop takes the form of a truncated sphere of radius r, the centres of the spheres being separated by a distance $2l$ (full-line diagram in Fig. 1).

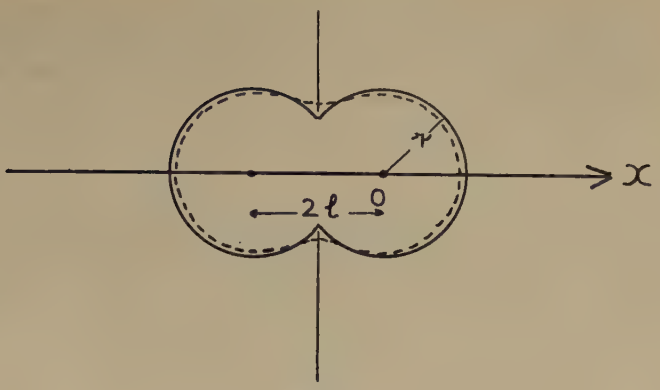

Figure 1.

Then the total volume V and total surface area S are given by

$$V = \tfrac{4}{3}\pi r^3 + 2\pi r^2 l - \tfrac{2}{3}\pi l^3$$

and

$$S = 4\pi r(l + r)$$

But during coalescence, V must remain constant, so that $dV/dl = 0$. This reduces to the condition $dr/dl = -(r - l)/2r$. Hence $dS/dl = 2\pi l(r + l)/r$, which is positive so long as l is positive. This shows that as the drops amalgamate (l decreasing) the total surface area will also decrease until a single spherical drop is formed..

In fact, the slope of the surface of the drop must be continuous everywhere (as shown by the pecked line in Fig. 1) but this will decrease the total area still further. The argument above will apply *a fortiori* to drops of unequal size.

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18 March 1957.

551.594.12 : 551.558.29

Charged nuclei and subsidence

By H. R. Byers

In a recent issue of the *Quarterly Journal* (**82**, pp. 428-455), R. C. Sagalyn and G. A. Faucher present their results of space and time variations of charged nuclei. In their Fig. 5, a comparison is made between the ' theoretical ' rate of decrease of total nuclei as determined by Smoluchowski's equation and the ' experimental ' rate of decrease obtained from their measurements. The authors point out that the observed rate of decrease is made up of the combined effects of coagulation, given by the equation, of inhomogeneities in the air mass and of subsidence.

In one of the situations which the authors presented in detail, subsidence was pronounced, as indicated by the series of temperature soundings obtained during the flight. Interest would have been added to the paper if the subsidence effects had been calculated. The following is an attempt at such a calculation.

The concept of the individual change (within a sinking parcel of air) and the local change (at a fixed level through which the parcels are passing) should be used. The changes in volume of the parcels as well as the changes in nucleus concentration must be considered, since the latter is given per unit volume. We have the well-known relation

$$\frac{d}{dt}(\alpha Z) = \frac{\partial}{\partial t}(\alpha Z) + w\frac{\partial}{\partial z}(\alpha Z) \quad . \quad . \quad . \quad . \quad . \quad (1)$$

where the total derivative represents the individual change and the partial derivatives the local change. Z is the total number of charged and uncharged nuclei per unit volume and α is the specific volume of the air. The velocity w is in the z axis, positive upward. In expanded form, the derivatives become

$$\alpha\frac{dZ}{dt} + Z\frac{d\alpha}{dt} = \alpha\frac{\partial Z}{\partial t} + Z\frac{\partial \alpha}{\partial t} + w\left(\alpha\frac{\partial Z}{\partial z} + Z\frac{\partial \alpha}{\partial z}\right) \quad . \quad . \quad (2)$$

The sinking parcel changes its volume adiabatically, so

$$\frac{d\alpha}{dt} = -\frac{c_v}{p}\frac{dT}{dt} = -\frac{c_v}{p}\frac{dT}{dz}\frac{dz}{dt} = \frac{c_v}{p}10^{-4}w \quad . \quad . \quad . \quad (3)$$

Substituting this equality in (2) and solving for w, we obtain

$$w = \frac{\alpha(\partial Z/\partial t - dZ/dt) + Z(\partial\alpha/\partial t)}{10^{-4}Z c_v/p - \alpha(\partial Z/\partial z) - Z(\partial\alpha/\partial z)} \quad . \quad . \quad . \quad (4)$$

The situation considered is that of 13-14 May 1954. The change of Z in the sinking parcel (dZ/dt) is given by Smoluchowski's equation, represented in the authors' Fig. 5. The measured or 'experimental' quantities in Fig. 5 and data of Fig. 3 can be used to determine the local changes. Taking values for the period 2245 hr to 0245 hr, 13-14 May, at 3,000 ft, we obtain the following: $\alpha = 897$ cm^3 gm^{-1} (mean); $\partial Z/\partial t - dZ/dt = -175$ cm^{-3} hr^{-1} $= -4{\cdot}87 \times 10^{-2}$ cm^{-3} sec^{-1}; $Z = 6{,}000$ cm^{-3} (mean); $10^{-4} Z c_v/p = 4{\cdot}94$ cm^{-1} gm^{-1}; $\partial\alpha/\partial t = 1{\cdot}7$ cm^3 gm^{-1} hr^{-1} $= 4{\cdot}74 \times 10^{-4}$ cm^3 gm^{-1} sec^{-1}; $\partial Z/\partial z = -600$ cm^{-3} per 1,000 ft (for the layer 3,000 to 4,000 ft) $= -1{\cdot}97 \times 10^{-2}$ cm^{-4}; $\partial\alpha/\partial z = 13$ cm^3 gm^{-1} per 1,000 ft $= 4{\cdot}26 \times 10^{-4}$ cm^2 gm^{-1}.

These values in Eq. (4) produce

$$w = \frac{-43{\cdot}6 + 2{\cdot}84}{4{\cdot}94 + 17{\cdot}7 - 2{\cdot}56} = -2{\cdot}03 \text{ cm sec}^{-1} = -240 \text{ ft hr}^{-1}$$

This rate of subsidence seems to be too high. The base of a strong temperature inversion, according to the authors' Fig. 3 (c), sank during this period from 8,000 ft to 6,900 ft in approximately 4 hr 40 min, or, roughly, -235 ft hr^{-1}. Since w must be zero at the ground, it should be a smaller negative quantity at 3,000 ft than at the higher altitude, perhaps less than 200 ft hr^{-1}.

The numbers used here were interpolated over finite time and height intervals from the author's graphs and no doubt could be improved upon from the original data. It is to be noted that the first term in the numerator and the second term in the denominator are the really important ones in this case. They depend on the time and space changes of nuclei which varied during the flights, making the assignment of mean values difficult. The dZ/dt from Smoluchowski's equation is highly sensitive to particle size, which also varied.

It is reasonable to conclude that, allowing for inhomogeneities in the air mass, subsidence rates computed from Eq. (4) can be expected to account for the differences between theoretical and observed changes in nuclei concentrations. In the case of 13-14 May 1954, the subsidence rate was fairly high. The Washington weather map for the period shows that the flight area was in the centre of an anticyclone.

The University of Chicago,
Chicago 37, Illinois, U.S.A.
22 April 1957.

551.577.33 : 551.590.24 : 523.51

Meteor dust streams and daily precipitation

By H. Melzer

I should like to call the attention of your readers to four papers which I have published on this subject in Journals which they might not normally see : *Urania* of the Sociedad Astronómica de España y América, Tortosa Ap. 9, Spain, No. 244, July/September 1956; *Revista do Clube de Engenharia*, No. 230, October 1955 and No. 250, June 1957; *Aguas e Energia Eléctrica*, No. 29, March 1957, Rio de Janeiro. The first has an English text and the others have English summaries.

I have tabulated the daily rainfall for a number of stations distributed in all continents for as many years as are available (up to 119 years) and joined the absolute and four subsidiary monthly maxima by curves. I find that these curves approximate to conics and are almost exactly of the same shape for all stations. For some stations, however, there are slight shifts of all curves as a whole in the same direction relative to the corresponding curves of other stations. Generally on intersections greater rainfall figures are to be found, and the greater the number of intersecting curves, the greater the anomaly of rainfall.

In my papers I relate these curves to the intersection of the earth's orbit with shells of the finest concentrated meteor dust responsible for zodiacal light and accompanying certain meteor streams. These relations raise the possibility of long-term prediction of rainfall amounts for any pluviometric station in the world.

Icarai Palace Hotel,
Praia das Flechas,
Niteroi-E. do Rio-Brazil.
9 March 1957.

551.510.534 : 551.508.952

New methods in the ozone problem

(Discussion meeting held on 16 Jan. 1957. See this volume, p. 278)

The problem of atmospheric ozone has continuously interested spectroscopists and chemists because of the great problems in technique which are involved. Meteorologists have also been interested in the possibility of using ozone as an additional forecasting tool mainly owing to the work of Dobson in this country, but they are not normally familiar with the techniques and the possibilities of the various methods of observation. For this reason an afternoon and evening meeting was held by the Society on 16 January 1957 at which experts discussed various techniques which have developed mainly during the last ten years.

The discussion was introduced by the Chairman, Sir Charles Normand (Oxford), who emphasized that there are at present no reliable direct observations of the concentration in the 10 to 20 km layer where most of the meteorologically induced changes may be expected. New methods are therefore essential if the problem is to advance beyond its present stage.

The first speaker, Dr. R. M. Goody (Imperial College), discussed two papers published jointly with colleagues in the *Quarterly Journal* (*Q.J.*, **82**, pp. 177, 217). He showed that the comparatively neglected field of infra-red spectroscopy could yield information on the vertical distribution of ozone by means of measurements on the 9·6 μ band. Dr. Goody emphasized that the only reliable quantity which could be obtained in this country was the total amount in the troposphere, but suggested the method was valuable because measurements could be made regularly and with little expense from a ground station.

Dr. H. K. Paetzold (Max Planck Institute, Weissenau) then described a filter radiosonde. This is an attempt to make the spectrographic method of Regener into a practical routine method and, judging by the comparison of the speaker's results with more accurate methods, a considerable degree of success had been achieved. The radiosonde will be used on a routine basis throughout the International Geophysical Year.

Dr. N. A. Gebbie (Liége University) had been asked to make an appreciation of the possibility of using microwave techniques. His note was read by Dr. C. D. Walshaw, but it led to the disappointing conclusion that this field did not present many opportunities.

Dr. Paetzold then described his attempts to improve the method of measuring ozone temperatures from the contrast in the Huggins bands. This work can be looked upon either as a way of measuring a stratospheric temperature or of finding the mean height of the ozone, given the temperature distribution. The method seems to be accurate enough for either purpose, but it has yet to be applied.

A clear and concise review of the principles of chemical ozone measurements was given by Dr. E. Glueckauf (Harwell). A list of his references is appended since much of the work is in specialist journals not normally accessible to the meteorologist. Although he was not attempting to put forward original ideas, Dr. Glueckauf did suggest a method of concentration of ozone which might be of use.

The most unusual paper of the meeting was by Dr. A. W. Brewer (Oxford) who described the progress of work in his laboratory on a chemical radiosonde. The instrument depended upon the classical potassium-iodide reaction and the current required to dispose of the iodine formed was measured. This is an absolute method which can be operated continuously and can measure a concentration of one part in 10^9 in about 20 sec. Ingenious design had been required to make the instrument light enough and reliable enough for balloon work. It was clear that members of the audience were greatly impressed by Dr. Brewer's work and it is to be hoped that the radiosondes will be in service in the near future.

The final paper by Dr. Paetzold described the method of finding ozone distribution from a lunar eclipse. This is a very interesting problem which has two real advantages over other techniques. It is possible to make simultaneous measurements over a wide range of latitudes and concentrations can be measured to higher levels than is possible by other methods.

The audience had mainly come to listen and did not contribute much to the discussion, but many thought the meeting of great interest, particularly because of the extraordinarily wide range of techniques described.

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R.M.G.

551.511 : 551.509.317 : 681.14

The use of a stream function in a two-parameter model of the atmosphere

By F. H. BUSHBY and VERA M. HUCKLE

(Read 20 February 1957. See *Q.J.*, **82**, p. 409)

Dr. R. S. Scorer : I would like to ask two questions: (1) would it not be illuminating to display the result of subtracting the actual from the forecast chart; the density of lines representing the inaccuracy of the forecast ? (2) Has Mr. Bushby got cases in which the air is rotating in space in the opposite direction to the earth ?

Dr. B. J. Mason : I should like to ask whether replacement of the contour chart parameter by the stream function appreciably increases the length of the computation, and whether this introduces extra difficulty or inconvenience in deriving the 1,000-mb contour chart.

Dr. R. C. Sutcliffe (*President*) : May I put two questions ? First, I understand the stream function is intended to give a representation of the actual wind more accurate than that given by the geostrophic wind. Does it in fact do so ? Has this been checked, or could it be checked by direct comparison with measured winds of which we have good records in the British Isles ?

My second question concerns the mixture of a stream function and thickness contours. How is the 1,000-mb contour field obtained from these two parameters ?

Mr. F. H. Bushby (*in reply*) : In reply to Dr. Scorer I agree that the differences between actual and forecast charts produce sets of isopleths which are of interest, but we decided not to reproduce these sets in the paper as they did not seem essential.

In three or four hand computations we have found anticyclones in the 500-mb pattern for which the finite difference approximations to the geostrophic vorticity imply that the air is probably rotating in space in the opposite direction to the earth.

The replacement of the contour-height variable by the stream function increases the time of a 24-hr time integration by about 15 per cent. The time to re-convert the forecast stream function back into a contour field is quite small (only a linear partial differential equation has to be solved) and hence the forecast 1,000-mb contour chart can be obtained quite easily by subtracting the computed thickness chart from that contour field.

I agree with Dr. Sutcliffe that we should check that the stream function does provide a better representation of the wind field than the geostrophic approximation. However, I think it better to wait until we have computed more cases and then carry out a statistical check at the upper-air reporting stations in the British Isles.

551.524.32 : 517.512.

The representation of the annual temperature variation over central and Northern Europe by a two-term harmonic form

By J. M. CRADDOCK

(Read 20 February 1957. See *Q.J.*, **82**, p. 275)

Prof. P. A. SHEPPARD : Does a two-term series provide as good a fit to the annual variation of air temperature over the oceans, or at a small oceanic island, as it does for a land station ? The relatively high value of the ratio of the second to the first harmonic over the oceans suggests the fit might not be quite so good there.

Would Mr. Craddock consider including in the printed discussion a diagram showing the standard deviation of 5-day mean temperature, in addition to the mean, for a couple of stations which have noticeable singularities ? We should then see at a glance the practical and statistical significance of the singularities.

(*Added later*) : The spatial patterns of temperature anomalies shown by Mr. Craddock throw up a nice problem. Temperature anomalies, as scalar quantities, are convenient for representation and surface temperature is of immediate significance in weather. Moreover, long records of surface temperature are available. But, looking rather far ahead, one would hardly choose surface temperature as a fundamental parameter for a theoretical study of perturbations of the general circulation (or however one chooses to frame the theoretical problem of long-range forecasting). One might, I suppose, choose the 1,000 mb-500 mb thickness anomaly which will be closely correlated with the surface-temperature anomaly. Does Mr. Caddock have any indication so far of the dynamically and thermodynamically important feature of the atmosphere which is reflected in his patterns of surface temperature anomaly ?

Mr. L. C. W. BONACINA : I should like to say that the particular singularity, namely the warm spell in late May and early June, which Mr. Craddock in showing his slides realistically called ' our old friend,' is very much alive in the experience of an ancient person like myself. In southern England the tendency is distinctly noticeable for a hot thundery spell to mark the passing of May into June, later to give way to chilly northerly winds with grey skies which on Mr. Craddock's own monsoonal explanation represent a ready influx from the cold northern seas. It is pleasant sometimes to be able from vivid personal weather memories to resuscitate the dead bones of statistics.

Dr. R. C. SUTCLIFFE (*President*) : Although the singularity of 1°-2°C in early summer has an amplitude which is small compared with the variations from season to season it may have some prediction value. The interruption in the rising trend of temperature, with even a small decrease of mean temperature, accompanying the so-called European monsoon, is usually thought of as an advection effect but changes in cloudiness, rainfall and evaporation may be very significant in the energy balance.

Dr. D. M. GATES : With regard to the monsoon shown in June for some of the European continental stations, should this same type of monsoon not be noticed in mid-continent America if it is primarily a continental phenomenon ?

Dr. G. D. ROBINSON : Could Mr. Craddock tell us anything about the distribution of deviations from the value given by his two-term series. He showed an example of a very large anomaly over much of W. Europe in February 1956. How many normal years will be required before this period in February ceases to appear as a ' singularity ' in this type of analysis ?

Mr. J. S. CRADDOCK (*in reply, partly communicated*) : Prof. Sheppard raises three points :

1. I have not been able to find suitable data for an oceanic island on which to test the goodness of fit of the two-term harmonic form. However, in Fig. 2 of the main paper the percentage residual variances for Edinburgh and Glasgow, which have more maritime climates than the rest of the 41 stations examined, are not higher than would be expected, taking account of the number of years over which the data are averaged.

2. Fig. 1 shows graphs of the standard deviation of 5-day mean temperatures for four stations. These for Budapest were published by Peczely (1955) and are computed from the sums of squares. These for Kew, Potsdam and Uppsala were found by the quicker but less accurate ' range ' method; the range in each 5-day period in each decade was averaged over all decades for which data were available, and the mean divided by 3·08. The method is described by, *inter alios*, Brownlee (1949).

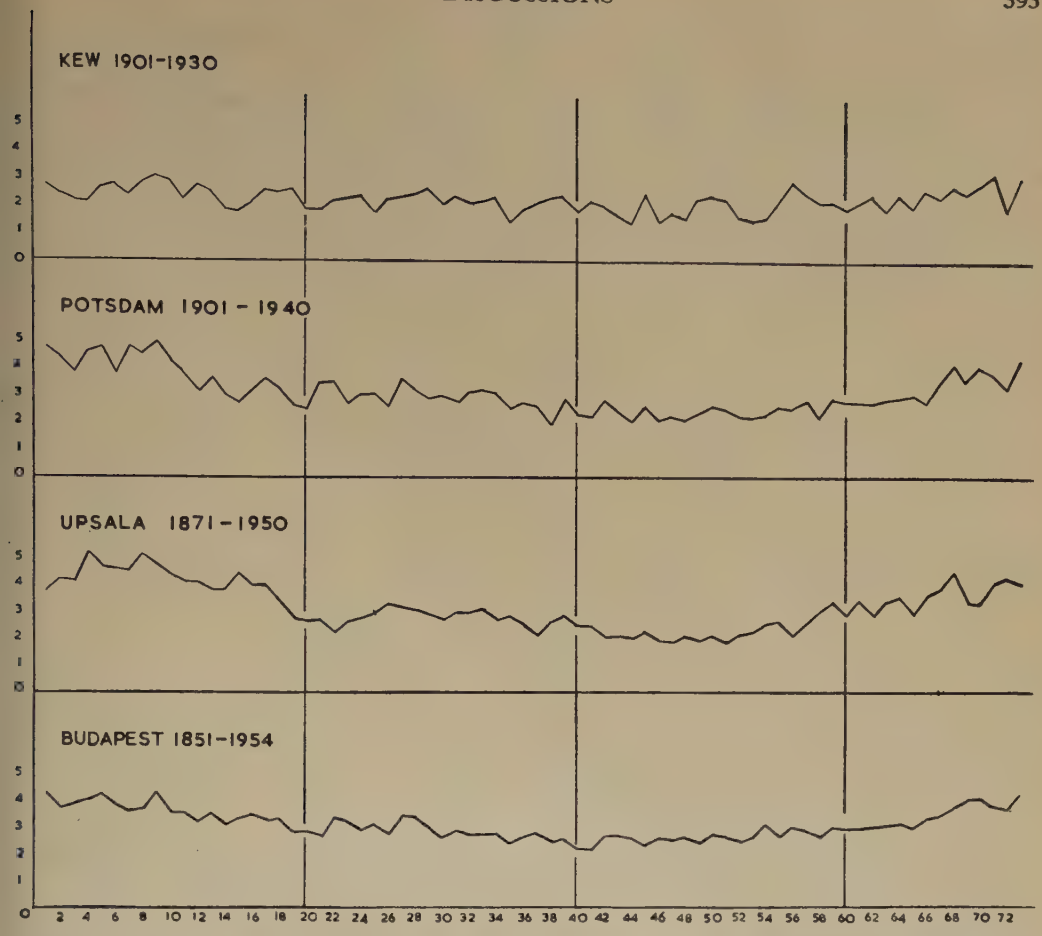

Figure 1. Standard deviations (°C) of 5-day mean temperatures for 4 stations. Abscissa, number of the 5-day period, starting from 1-5 Jan. as 1.

As regards the statistical significance of the singularities, say, at Potsdam, the level of the standard deviation given in Fig. 1 suggests that in late spring the standard deviation of a 50 yr mean should be about 0·4°C. The deviation actually occurring in the period 26-30 May is about + 1·0°C, a value likely to arise in a period chosen at random on only one in a hundred trials. Unfortunately the period 26-30 May was chosen not at random, but because it showed the largest deviation of any of the 73 periods covering the year and the station, Potsdam, was chosen, not at random, but because Fig. 5 of the main paper shows that the deviations at Potsdam are more impressive than those at other stations such as Glasgow or Tilsit. The net result of these two forms of selection, spatial and temporal, is that the odds against chance occurrence of a 1·0C° deviation are very much less than 100 to 1.

3. As Prof. Sheppard says, the use of surface-temperature anomalies was dictated by reasons of practical convenience, and is hardly the parameter one would choose for studying perturbations of the general circulation. However, experience during the last two years has shown that monthly mean patterns of anomalies of the 1,000-500 mb thickness and those of the air temperature at the earth's surface are very similar, and this similarity must extend to patterns of the anomaly of 500 mb contour height. Research is proceeding, but is impeded by the shortness of the period for which a hemisphere-wide network of upper-air observing stations has been in existence.

In reply to Mr. Bonacina, I share his pleasure in finding that a statistical feature is supported by personal observation. I feel, however, that if he is right in speaking of ' the dead bones of statistics ' then the fault lies, not with statistics but with statisticians.

In reply to Dr. Sutcliffe, although the early summer singularity shows up in the average, it may not be recognisable in individual years. For example, in 1955 the sequence of events was quite different. The question whether the ' monsoonal ' temperature fall is due to advection or

increased cloudiness, rainfall or evaporation could be settled by examining the individual years, but I have not had time to do this.

In reply to Dr. Gates I could not extend the investigation to America as I could not find any suitable data. I think the question will have to be left to some investigator who has ready access to the Punched Card Library of the United States Weather Bureau.

In reply to Dr. Robinson, my anwers to his questions are mostly contained in my reply to Prof. Sheppard. The negative anomaly in February 1956 could be cancelled by two or three large positive anomalies in other years, but I agree with him that there is a possibility that some of the ' singularities ' are no more than the residue left from the superposition of a few well-marked patterns which in most places tend to cancel each other out.

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551.577.11 : 551.574.1 : 551.501.81

Radar observations of showers suggesting a coalescence mechanism

By P. J. FETERIS and B. J. MASON

(Read 20 March 1957. See *Q.J.*, **82**, p. 446)

Mr. W. G. Harper : I should like to pay tribute to the ingenuity of Dr. Mason and Mr. Feteris in rebuilding their radar to scan just that area of the sky which is unknown territory to our 20° maximum elevation RHI radars, and in this way taking the greatest advantage of the limited power of their equipment.

From experience at East Hill I do not think many of the echoes reported in Table 1 were typical of summer showers in south-east England. A surprising number failed to grow at all after detection of the initial echo. This is I think most unusual. With the power output 500 kw and pulse-length 1·9 μsec of the East Hill 10-cm RHI these must have been very light showers. The summer of 1955 was indeed exceptionally dry.

A point not brought out in the paper is that a low-power radar looking almost vertically (for which each part of the shower column has a different slant range) will not record the same height of initial echo as a more distant high-power radar used at a low angle of elevation (for which the slant range has effectively a single value over the whole shower column). Will this not bias the comparison with temperature levels ?

Dr. B. J. Mason (*in reply, on behalf of both authors*) : I am inclined to agree with Mr. Harper that the summer of 1955 may have been rather unusual in regard to the incidence of shallow, warm showers which failed to grow appreciably after detection of the initial echoes. Certainly warm showers appeared to be much less frequent during the summer of 1956. The effect of range attenuation might, of course, be important for a shower directly overhead. But for a shower echo at range 5 mi with base at, say, 1 km and top at 3 km, the range attenuation of top relative to base amounts to only 1 db; the effect was therefore not important in the cases which we studied.

551.574.1 : 551.577.11

An inherent precipitation mechanism in cumulus clouds

By T. W. R. EAST

(Read 20 March 1957. See *Q.J.*, **83**, p. 61)

Dr. B. J. Mason : Assuming that droplets of radius up to 18 μ exist in a small cumulus at a few hundred metres above cloud base, Dr. East shows that if they are lifted at 1-3 m/sec along a saturated adiabatic, these larger droplets may grow by continued condensation and

coalescence to radius 50 μ after a further 20-30 min. But if we are to explain how the coalescence mechanism is initiated, it seems important to enquire into the origin of the 18 μ droplets from which Dr. East starts his calculations and to examine the time required for their formation.

To do this, it is necessary to consider the growth of a population of nuclei in rising cloudy air and to calculate the supersaturation and droplet radii as functions of time, in terms of the updraught speed, the temperature, pressure, and the size and concentration of the nuclei. Mr. Ghosh and I have recently made calculations of this kind along the lines of earlier work by Howell. To take a relatively simple example : if droplets in a concentration of 300/cm^3 are assumed to grow upon salt nuclei of mass 10^{-15} g while being carried up at 1 m/sec from cloud base at 10°C, 900 mb; after 600 sec their radii would be 9·6 μ and the supersaturation 0·07 per cent. Rare droplets of radius 18 μ could appear at this stage, 600 m above cloud base, only if they were formed upon nuclei containing at least 10^{-10} g of salt. This suggests that Dr. East has not really demonstrated that the coalescence mechanism can operate in shower clouds without the presence of giant nuclei since they appear necessary to provide the larger droplets which form the basis of his calculations.

The problem is also rather more difficult than Dr. East's paper suggests in that large droplets are found in surprisingly high concentrations in shallow clouds, 3,000-5,000 ft deep, and in which the average liquid-water content is considerably less than the adiabatic value and much less than the values contemplated by Dr. East. For example, the Meteorological Research Flight find droplets of radius 30 μ in concentrations of 10-100/l and 50 μ radius droplets in concentrations of order 1/l near the bases of such shallow clouds with water contents of only a few tenths g/m^3. Similar observations have been made in the United States by Weickmann and others.

In an attempt to explain how these large droplets may be formed, I have calculated the rate of droplet growth by condensation and coalescence on giant hygroscopic nuclei, entering the cloud base as droplets of saturated solution, as they are carried up at 1 m/sec among a population of 300 droplets per cm^3 growing on much smaller nuclei. The supersaturation and liquid-water content are evaluated as functions of time assuming the bubble core to rise undiluted. Computation is stopped after 1,500 sec when the cloud depth reaches 1,500 m. Thereafter, although new bubbles may continue to sustain the cloud, no further vertical growth is assumed. It is assumed that some of the larger droplets formed in a rising bubble later become transported to lower levels in the cloud by turbulent mixing, growing by coalescence on the way. The air carrying the droplets down will become warmed, and will remain saturated only by evaporation of some of the contained water. Its liquid-water content will therefore be considerably less than the adiabatic value for the level in question and will be largely determined by the degree of mixing with surrounding cloudy air.

My calculations show that a salt-nucleus of mass 10^{-10} g entering the base of the model cloud described above would attain a radius of 27 μ after 1,500 sec; thereafter, on being transported to lower levels through a cloud of average liquid-water content 0·5 g/m^3 and continuing to grow by coalescence, it would reach a radius of 50 μ in a total time of rather less than one hour. In the same time and under the same conditions, a nucleus of mass 10^{-11} g would produce a droplet exceeding 30 μ radius.

Thus a cloud only 1,500 m deep and lasting for about an hour, could produce droplets of radius 30-50 μ if it were provided with nuclei of mass 10^{-11} to 10^{-10} g. Over the oceans, in areas of spray formation, nuclei of 10^{-10} g are found in concentrations of about 1/l those of 10^{-11} g being about 10 times more abundant. Such concentrations of giant salt-nuclei are barely sufficient to explain the observed concentrations of large droplets in some small cumuli, bearing in mind that the numbers of nascent droplets carried up in ascending bubbles will become diluted as they become mixed downwards in the cloud. Over southern England the numbers of giant hygroscopic nuclei may sometimes be comparable to those measured over the oceans, but it would seem highly desirable to measure simultaneously from aircraft the concentrations of both the large cloud droplets and the giant nuclei, in order to determine whether the latter may account for the former.

Mr. W. G. Harper : I find that the growth curves of heavy shower echoes which have been recorded at the Radar Research Station of the Meteorological Office at East Hill show an interesting similarity to Dr. East's Fig. 8. The first echoes had an average depth of 11,000 ft. for those whose tops rose to above 30,000 ft. On an average their echo tops rose at 1,000 ft/min, but at the same time the echo bases fell at a slightly greater rate. Slight extrapolation of the growth curves back in time suggests that in our admittedly rather few cases the level at which the first detectable drops appeared was near or rather above the level of the 0°C isotherm (overall temperature range 0°C to − 9°C).

Mr. L. C. W. Bonacina : As regards the coalescence process, which seems to be returning to its own, I wish to raise a question. Rain and mist seen driving along a hill-side present to the eye a complex spiral pattern due, presumably, to turbulent eddies. What part do such eddies, in the light of what was said in one of the papers, play in coalescence ? What is the actual physical motion of the mist and rain particles producing the visual spiral pattern mentioned ?

As to the precipitation mechanism in cumulus clouds how close is the connection between the formation of cirrus at the top of a cloud and the discharge of rain below ? I gather from what has been said that when a cumulus frays out into cirrus rain will always fall, but that a cumulus could precipitate rain without developing cirrus.

Dr. T. W. R. East (*in reply*) : The entire precipitation process in cumulus from saturated clear air to rainfall at the ground involves a number of physical processes whose character and relative importance varies during the history of a raindrop. The common practice of dividing the process into three or four stages for purposes of study is, I feel, justifiable on the grounds that our understanding will improve more rapidly that way, though the transition from one stage to another may well be a gradual one (the start and finish of each stage must, of course, be checked against observation wherever possible). I chose to concentrate on the part of the process in which droplets of about 18 μ radius, hitherto considered too small to grow appreciably by coalescence, do in fact reach drizzle-drop size in a few minutes. Dr. Mason chooses to study an earlier stage, and I am looking forward to seeing his results in print. I was not dependent on any theoretical treatment of the first stage since reliable observations were available for my starting point.

(*Communicated*). In comparison to the population of cloud droplets, the number of raindrops is very small, so that admittedly they could originally have consisted of droplets grown to 30 μ radius by condensation, present in immature clouds or near cloud base, but rarely detected because of their relative scarcity. Such an hypothesis, however, cannot account for the large numbers of droplets of 50 μ and over (of the order 1,000 per litre) detected in dense cumulus congestus and in cumulonimbus. My main purpose has been to show that they can be produced from the known populations of droplets of under 20 μ. Once this *inherent* capability of cumulus clouds is accepted, it is still open to argument whether in some circumstances the rare 30 μ droplets found among under-20 μ populations, arising from giant nuclei, provide an alternative source of raindrops. My contention is that the mere fact of warm precipitation does not imply that these 30 μ droplets were present on all such occasions. Where 18 μ and 10 μ droplets are found in the same fresh cloud parcel, the larger droplets were, of course, condensed on to larger hygroscopic nuclei, but I had thought that the use of the term ' giant nuclei ' intended something rather more significant.

Mr. Harper's observations of the initial appearance and growth of echoes in cumulus are particularly valuable, and should take their place alongside those of Battan and others for North America and Bowen for Australia.

Eddies such as Mr. Bonacina reports would have the effect of driving larger droplets past the smaller ones by centrifugal force, so that a few collisions might be caused in addition to those caused by gravity. However, the radial acceleration in these curved paths would be much less than that due to gravity and would increase the rate of growth by collision by only a negligible amount. Presumably, if the air were to travel in a circular path the mist droplets would spiral outwards slightly, and raindrops more markedly. According to the modern theory of turbulence, however, in cumulus, an eddy is unlikely to persist for more than a quarter turn. In practice, the motions would be three-dimensional and observations might be rather difficult to interpret.

The formation of cirrus is usually taken to signal the presence of sufficient ice to produce rain by the Bergeron process. McCready, Smith, Todd and Beesmer (' Physical evaluation of cloud seeding effects,' 1956. Study prepared for the Advisory Committee on Weather Control by Meteorology Research, Inc., Pasadena, California) used the numerical results of our work, together with computations of their own, to calculate the time of fall-out of rain from cloud base in relation to observed rate of rise of cloud top. They applied this method to 14 cumulus clouds at Boca Raton, Florida. The actual time of fall-out was, on the average, 2½ min later than that calculated for the all-water mechanism (the standard deviation on this time difference was 2·3 min). Similar calculations based on an ice-crystal and accretion mechanism showed a calculated fall-out time averaging 14½ min *after* the observed fall-out time (standard deviation 8 min). Clearly the first precipitation to reach the ground was not connected with the ice mechanism in these examples.

Report of the Council
for the year 1956

Submitted to the Annual General Meeting, 17 April 1957

General

The Council are pleased to report again a year of satisfactory progress. For the sixth year in succession the total membership has shown a small increase over the previous year.

As with other Societies, the continuing rise in costs, together with other factors explained in the Finance section, have had an adverse effect on the Society's finance, and this matter has been receiving very careful consideration of Council. It is hoped to increase the yield from the Society's investments by investing a proportion of the funds in non-trustee stocks, as explained in the Finance section.

The third two-day Summer Meeting was held at Aberystwyth. Details are given later in the Report.

Two vacation courses were held in conjunction with the Field Studies Council, one, mainly for undergraduates, at the Malham Tarn Field Centre and the other for sixth-form pupils at the Flatford Mill Field Centre. Both courses were very successful; 16 students were subsequently elected to membership of the Society.

Council

From 1 January to 30 September the constitution of the Council was as recorded in the 1955 Annual Report.

From 1 October 1956 the following Council held office :

President
Reginald Cockcroft Sutcliffe, O.B.E., Ph.D.

Vice-Presidents
Professor Gordon Manley, M.A., M.Sc.
Sir Oliver Graham Sutton, C.B.E., D.Sc., F.R.S.
Richard George Veryard, B.Sc.
Thomas Wilson Wormell, M.A., Ph.D.

Vice-President for Canada
Donald Patrick McIntyre, Ph.D.

Vice-President for Scotland
James Robert Atkinson, M.A.

Treasurer
James Strachan Farquharson, M.A., D.Sc.

Secretaries
Arthur George Forsdyke, Ph.D., D.I.C., A.R.C.S. (Librarian)
Patrick John Meade, O.B.E., B.Sc.
Howard Latimer Penman, M.Sc., Ph.D., F.Inst.P.
George David Robinson, Ph.D., F.Inst.P. (Editor)

Councillors
Charles James Boyden, B.A.
James Marston Craddock, M.A.
Instr. Capt. Charles Roy Darlington, B.Sc., R.N.
Thomas Vivian Davies, M.Sc.
Thomas Herbert Ellison, M.A., Ph.D.
Ronald Westlake Gloyne, B.Sc.
Robert Frederick Jones, B.A.
Hubert Horace Lamb, M.A.
Basil John Mason, D.Sc.
Frank Pasquill, D.Sc.
John Stanley Sawyer, M.A.
Charles Edward Wallington, M.Sc.

The Council met nine times during the year and there were thirty-nine Committee meetings.

Meetings

Eight Ordinary meetings and two Special General Meetings were held in the Society's rooms during the year, including a joint discussion in January with the British Glaciological Society on 'Snow accumulation and ablation.' A joint two-day conference with the Physical Society on 'Cloud physics' was also held on 4-5 January at the Department of Meteorology, Imperial College.

Mr. N. Bradbury gave a Popular Lecture on 'Aviation meteorology' in November and December. The Society's series of Popular Lectures was extended for the first time to the Scottish and Manchester Centres.

The Annual General Meeting was held in the Society's rooms on 25 April when the Report of the Council and the Annual Accounts for 1955 were adopted. The Presidential Address was given by Dr. R. C. Sutcliffe, O.B.E., his subject being 'The moisture balance of the atmosphere'.

The Symons Memorial Lecture was given on 11 April by Mr. J. Paton on 'The Polar aurora'.

The Summer visit on 11 July was to the British Scientific Instruments Research Association (S.I.R.A.), Elmstead Woods, Kent, by kind invitation of the Director. Council wish to express to him and to Dr. A. J. Maddock, who was responsible for the arrangements, their appreciation of a most enjoyable visit.

Two-day Summer Meeting, Aberystwyth, 26-28 July 1956

The third two-day Summer Meeting arranged by the Society was held in honour of Sir David Brunt, Sec. R.S., whose seventieth birthday fell that year, and by kind invitation of the Vice-Principal, Professor R. M. Davies, took place at the Pantycelyn Hall of Residence of the University College of Wales, Aberystwyth.

The meeting consisted of two scientific symposia with an attendance at each of about 60 members and guests. The first symposium was opened by the reading of papers on 'An ionization method of measuring air density' and was followed by a discussion on 'Topographical influences on airflow'.

The second symposium consisted of two papers dealing with local climatology, followed by a discussion on 'Orographic effects in relation to rainfall'.

A most enjoyable excursion was arranged, under the leadership of Professor E. G. Bowen (Aberystwyth), to the reservoirs of the Claerwen and Elan valleys, by kind invitation of Mr. C. A. Risbridger, B.Sc., M.I.C.E., City of Birmingham Water Engineer.

A visit was also paid to the National Library of Wales, Aberystwyth, by kind invitation of the National Librarian, Dr. Thomas Parry.

Council wish to record their sincere thanks to Professor E. G. Bowen, who was responsible for the local arrangements.

Publications

The inflow of scientific communications for the *Quarterly Journal* was rather less than in previous years. Fifty-seven papers were submitted for publication during 1956 as against sixty-nine in 1955. Thirty-five of the papers have already been accepted for publication. The average time between receipt of approved copy and publication in the *Journal* was about four months. The four numbers of the 1956 *Journal* together contained 559 pages, as compared with 662 pages in 1955.

Council are greatly indebted to the Editor, Dr. G. D. Robinson, the Editing Committee and the proof readers, Mr. E. L. Hawke and Cdr. C. R. Burgess, for their services in the production of the *Journal*.

Thanks to the enthusiasm of its Editing Committee, *Weather* successfully completed its tenth volume and marked the occasion by announcing an Essay Competition.

Membership

Table A

	Fellows and Foreign Members			Associates	Students	Honorary Members	Total
	Annual	Life	Total				
Home Members							
31 Dec. 1955	1,392	174	1,566	28	107	17	1,718
Since elected	99	—	99	8	19	—	126
Re-admitted	3	—	3	—	—	—	3
Transferred	12	2	14	—	− 14	—	—
Deceased	− 8	− 3	− 11	—	—	− 1	− 12
Resigned	− 33	—	− 33	− 1	− 4	—	− 38
Lapsed	− 48	—	− 48	− 1	− 7	—	− 56
31 Dec. 1956	1,417	173	1,590	34	101	16	1,741
Canadian Branch							
31 Dec. 1955	46	6	52	298	—	1	351
31 Dec. 1956	58	5	63	272	—	—	335
Grand Total							
31 Dec. 1955	1,438	180	1,618	326	107	18	2,069
31 Dec. 1956	1,475	178	1,653	306	101	16	2,076

Deaths

The Council record with sorrow the death of Dr. T. Okada, Honorary Member, elected in 1925, and Symons Medallist (1924); and of Prof. J. Patterson, Honorary Member, elected in 1941 (Fellow 1920).

The following deaths are also recorded with regret, the dates of election being appended to the names : H. W. Apperley (1930), S. E. Ashmore (1932), H. C. Busk (1910), A. Hamilton Rice (Life Foreign Member, 1931), A. W. Lee (Life Fellow, 1923), A. G. Maddock (1950), I. C. Roberts (1917), W. Smith (1924), G. Swoboda (Foreign Member, 1953), Sir Arthur G. Tansley (1938) and T. L. A. Waite (1941).

Library

Increasing use was made of the Library, both for loans and reference purposes, and a substantial number of enquiries was dealt with.

To assist further in the preservation of the older valuable books, glazing has been provided for the shelves at the rear part of the Lecture Room. A considerable number of the older books which are valuable assets of the Society were rebound.

In order to safeguard the negatives of the Clarke and Cave collections of cloud photographs against loss and damage two sets of full-plate prints are to be made. These would be suitable, if required, for the production of negatives. A selection of 150 negatives has been made from the collections for this purpose. In addition 35 mm negatives are to be made for use in the production of lantern slides or film strips.

During the year 16 new series of journals or memoirs have been received from countries which already receive the *Quarterly Journal* in exchange. A former journal under the new title *Beiträge zur Physik der Atmosphäre* has recommenced publication and is received on exchange.

Appreciation has been expressed by several of the more important libraries of receipt from the Society of the catalogue of journals, memoirs, technical papers, etc., held in the Library.

Staff

At the close of the year the staff consisted of Mr. M. E. Shoenberg, B.A., Assistant Secretary; Mr. R. S. Read, I.S.O., M.A., B.Sc., A.R.C.S., F.Inst.P., Assistant Librarian; Mrs. A. Wallis, Accounting Assistant; Mrs. W. A. Steward, Publications Clerk and Editorial Assistant. Miss E. Crafts, *Weather* clerk, was promoted to this post from Junior Clerk as from 16 July, on the resignation of Miss G. C. Jones. Mrs. J. G. Johnston resigned her post as Secretarial Assistant and was replaced by Miss I. Dowsing and later by Miss M. L. Chidley. Miss E. Hooper replaced Miss E. Crafts as Junior Clerk on 7 August.

The Officers and Council have pleasure in recording their appreciation of the loyal and efficient service of the staff.

Branch and Centres

Canadian Branch

Eight meetings were held in Toronto during the year, in addition to the Annual Business Meeting in January. In October, the Second National Meeting of the Royal Meteorological Society, Canadian Branch, was held in Toronto, at which four papers were presented.

The Committee consisted of : President, R. C. Graham; Vice-President, W. L. Godson; Vice-President for Canada, D. P. McIntyre; Treasurer, C. Goodbrand; Secretary, R. Lee; Councillors-at-Large, Chairman of the Montreal Centre; Chairman of the Winnipeg Centre; H. Cameron, Ottawa; R. V. Tyner, Vancouver; L. B. Foster, Greenwood.

Montreal Centre. Eight meetings were held during the year, at which the average attendance was twenty-nine. During the week 11-16 June, members attended the Annual Meetings of the Royal Society of Canada and the Canadian Association of Physicists at the University of Montreal. The Officers were : Chairman, P. Johns; Secretary, T. W. R. East; Treasurer, M. Kwizak. After the transfer of P. Johns to Toronto, T. W. R. East and G. H. Washburn were elected Chairman and Secretary respectively.

Winnipeg Centre. Four meetings were held in 1956. The Officers were : Honorary Chairman D. M. Robertson; Chairman, S. V. A. Gordon; Vice-Chairman, D. S. McGeary; Secretary, M. Hardman; Treasurer, R. Holbrook.

Scottish Centre.

Four meetings were held during the year at Edinburgh University and an excursion was made in May to the meteorological establishment at Leuchars. In March, under the auspices of the Glasgow Corporation Education Committee, Mr. F. H. Ludlam delivered his Popular Lecture on ' Clouds and rain making'.

The Committee consisted of : Chairman and Vice-President for Scotland, Sir Ernest Wedderburn, succeeded in October by Mr. J. R. Atkinson; Secretary and Treasurer, Mr. J. Paton, succeeded in October by Mr. D. H. McIntosh; and nine other Fellows.

Manchester Centre

Six meetings were held during the year at the University of Manchester. In March, under the auspices of the Manchester Education Committee, Mr. P. J. Meade repeated, at the College of Technology, his Popular Lecture for sixth-form pupils on ' Fog and atmospheric pollution'.

The Committee consisted of : Chairman, Professor P. R. Crowe; Secretary, Dr. Ruth H. Rogers, followed by Mr. N. K. Horrocks in October; Treasurer, Mr. G. Reynolds; Assistant Secretary, Mr. N. K. Horrocks, followed by Dr. H. P. Palmer in October; and nine other Fellows.

The Centre is greatly indebted to Dr. Rogers for her services, which have contributed so much to the development of the Centre since its inauguration.

Representatives

The Society's representatives on other scientific bodies were :

Royal Society Committes
Gassiot Committee, President; *British National Committee for Geodesy and Geophysics*, Sir Graham Sutton (1952-57), T. W. Wormell (1955-60); *British National Committee for Physics*, F. Pasquill (1955-60).

Institute of Physics Committees
Journal of Scientific Instruments Advisory Committee, F. Pasquill (January-September), H. L. Penman (October-December); *British Journal of Applied Physics Advisory Committee*, G. D. Robinson.

Cambridge University Meteorological Committee
Sir Charles Normand.

Royal Astronomical Society
Geophysical Committee, G. D. Robinson.

Advisory Committee for the Meteorological Office, Edinburgh
Sir Charles Normand.

British Glaciological Society
Professor Gordon Manley (new representation agreed by Council 21 November 1956).

Awards

The following awards were presented at the Annual General Meeting on 25 April :
Buchan Prize, in respect of the period 1951-1955 : Frank Henry Ludlam, D.I.C., for his outstanding contributions to the study of cloud physics.

Hugh Robert Mill Medal and Prize for 1956 :

John Stanley Sawyer, M.A., for his outstanding contributions to the science of meteorology with particular reference to rainfall.

Darton Prizes for 1955

Home awards :
1st Prize : Mr. R. J. Murgatroyd.
2nd Prize : Mr. S. E. Ashmore.

Canadian awards :
1st Prize : Mr. Roy Lee.
2nd Prize : Mr. R. Anderson, Mr. B. W. Boville and Mr. D. E. McClellan. (jointly)

Napier Shaw Memorial Prize

The first award of this prize, instituted in 1954 to mark the centenary of the birth of Sir Napier Shaw, was awarded to Dr. Norman A. Phillips, the Institute for Advanced Study, Princeton, U.S.A., for his paper ' The general circulation of the atmosphere : a numerical experiment ' (*Q.J.*, **82**, p. 123). The award consisted of a prize of £100 and was announced at the June meeting, where Dr. Phillips was present to read his paper.

Scientific Activities Fund

During the year the following grants were made :

Oversea delegates, towards travelling expenses for attending the joint Conference with the Physical Society on ' Cloud physics ' held in January at the Department of Meteorology, Imperial College.

Mr. C. E. Wallington, in respect of attendance in July at the sixth Scientific Congress of O.S.T.I.V. at St. Yan, France.

Mr. A. J. Whiten, towards the cost of the calculating machine purchased by him for his meteorological work.

In addition the following grants for payment in 1957 were approved :

Prize for a *Weather* essay competition.

Field Studies Council towards the expenses of a pioneer meteorological vacation course on ' The weather and flight ' mainly for members of Flying and Gliding Clubs, to be held in July 1957 at the new Field Centre at Preston Montford, Nr. Shrewsbury.

The Scientific Activities Fund came into existence largely as a result of a generous donation from the Munitalp Foundation, U.S.A. and a report covering the first four years' working of the fund was accordingly sent to that organization.

British Rainfall Organization Fund

For some time past, 62 Camden Square, the property administered under the above fund, has been becoming increasingly derelict, and the situation of the observing station maintained there by the Society no longer satisfied the necessary meteorological standards.

At a special General meeting held on 19 December it was unanimously agreed that the property should be sold.

A plaque commemorating the long association of the British Rainfall Organization, and especially that of the late G. J. Symons, F.R.S., with 62 Camden Square, has been affixed to the property. A clause will be included in the Contract of Sale for the retention of the plaque at a specified position on the property.

By arrangement with the St. Pancras Borough Council the observing station is being transferred to a nearby site in the small park in Camden Square.

Bequests

The Society received in August £1,000 from a bequest by the late Mr. A. Hampton Brown, formerly Assistant Secretary to the Society, who retired in 1935 after 40 years' service.

It was the testator's desire, without imposing any trust, that the bequest should be used chiefly for the maintenance of the library of the Society.

The Council decided that up to £30 should be spent on preparing 35 mm negatives from selected prints in the Clarke and Cave Collections of cloud photographs. The remainder is being invested to form the ' Hampton Brown Fund ' ; in expending capital or interest, priority will be given to the needs of the Library.

Finance

It will be seen that there is a deficit this year of £253 as compared with one of £29 last year. The deficit for 1956 is due solely to the fact that the Authorities of the Inland Revenue decided to withhold from the Society the refund of income tax hitherto paid on covenanted subscriptions. They claim that the entitlement of the Society to this refund (£460) is on insecure foundations, now being tested in a case which is to go to the Court of Appeal. Until this case is settled and the claim of the Society to the refund re-established it would be unwise to include in the income of the Society any sum in respect of this item.

As a result of a Special General meeting on 25 April 1956, and of the Annual General Meeting following it, the wording of By-law 86 having been changed, it was agreed that a sum up to £7,000, of the Society's capital, could be invested in non-trustee stock during the ensuing twelve months. In fact the first investment of this kind was not made until early in 1957 so that it cannot appear in this account.

The Royal Society allocated to the Society a sum of £500 from the Government Fund for Grants-in-aid of Publications. An additional sum of £400 was also allocated for the Library. The receipt of these sums is acknowledged with gratitude.

The preceding sections of this report provide ample evidence of the health and vigour of the Society. Nevertheless it would be unwise to ignore the adverse balance of the Expenditure and Income Account which indicates that the Society, faced with continually rising costs, may find difficulty in maintaining all its varied activities at a level that members would desire. Council are therefore giving anxious thought to the financial problems confronting the Society and will make every effort to ensure that current expenditure is met out of income.

1955			£	£
£	£			
		ACCUMULATED FUND—		
		Excess of Assets over Liabilities at 1 Jan.		
12,516		1956	12,487	
		Deduct Transfer to Library Equipment		
—		Fund	508	
12,516			11,979	
29		*Deduct* Excess of Expenditure over Income	253	
12,487				11,726
—		Library Equipment Fund		508
		CURRENT LIABILITIES—		
1,181		Creditors and Accrued Charges—	710	
		Amounts received in advance—		
	454	Publications	683	
	330	Subscriptions	334	
784			1,017	
220		Due to Bankers	1,560	
2,185				3,287
£14,672				£15,521

We have examined the above Balance Sheet and annexed Income and Expenditure Account which in our opinion respectively give a true and fair view of the state of the Society's affairs as at 31 December 1956, and of its income and expenditure for the year ended on that date.

Spencer House,
South Place,
LONDON, E.C.2. BALL, BAKER & Co.
27 March 1957. *Chartered Accountants.*

1955						£	£
£	£						
		FIXED ASSETS—					
		Freehold Property – Estimated value (in 1925) of the Society's House, 49 Cromwell Road					5,375
5,375		Furniture and Equipment—					
			At 1.1.56	Additions during year	Accumulated Depreciation	Net	
	1,186	General	1,383	153	313	1,223	
	405	Library	457	148	97	508	
1,591			£1,840	£301	£410	£1,731	1,731
		(No amount has been included for the value of the books of the Library and the stock of publications)					
		INVESTMENTS—					
		Valued at market price 31.12.54—					
	924	£1,400 2½% Consolidated Stock ..				924	
	1,206	£1,250 4% Consolidated Stock 1957 or after				1,206	
	796	£900 3½% Conversion Loan 1961 or after				796	
	206	£200 4% Funding Stock 1960/90 ..				206	
	645	£649.1.7 Government of Australia 4% Registered Stock 1955/70				645	
	2,030	£2,000 Government of Australia 4% Registered Stock 1966/68				2,030	
5,807							5,807
		(Market Value 31.12.56 £4,726)					
		(Market Value 31.12.55 £5,132)					
		CURRENT ASSETS—					
50		Stock of Stationery					50
		Debtors—					
		£	*Proceedings Toronto Meteorological Conference* 1953 ..			£	
	49					54	
	345		Subscriptions			804	
	390		Publications			559	
	784					1,417	
	150		*Less* Provision for bad and doubtful debts ..			150	
	634					1,267	
	18	Payments in advance				50	
	1,197	Deposit at Post Office and Cash in Hand				1,241	
1,849							2,558
£14,672							£15,521

GENERAL

INCOME AND EXPENDITURE ACCOUNT

EXPENDITURE

1955 adjusted* £	£		£	£
		ESTABLISHMENT—		
	46	Architect's Fees	—	
	12	Water Rates	27	
	275	Cleaning Wages and Materials	270	
	24	Insurance	27	
	206	Lighting and Heating	211	
	705	Repairs and Maintenance	170	
1,376	108*	Depreciation of Furniture and Equipment	116	
				821
		ADMINISTRATION—		
	2,458	Salaries	2,645	
	349	Contribution to Staff Pension Fund	388	
	165	Printing	95	
	174	Stationery	160	
	271*	Postage, Carriage and Packing	274	
	85	Telephone	76	
		Audit and Accountancy Fees—		
	105	Royal Meteorological Society	105	
	31	*Weather* Magazine	31	
3,657	19	Sundries	32	
				3,806
		PUBLICATIONS, ETC.—		
	3,458*	*Quarterly Journal*	3,647	
	339*	Illustrations	315	
	377	Reprints of Papers	300	
	22	*Monthly Weather Report*	8	
	934	Members' copies of *Weather*	994	
5,147	17	Second-hand Journals	2	
				5,266
		LIBRARY—		
	445	Salary	486	
	164	Books, Bindings, etc.	192	
643	34*	Depreciation on Fixtures and Fittings	45	
				723
		MEETINGS' EXPENSES—		
	157	London	178	
	40	Scotland	40	
	106	Centenary Exhibition of Scottish Meteorological Society	—	
333	30	Manchester	30	
				248
63		PROVISION FOR BAD AND DOUBTFUL DEBTS—		28
		GRANTS AND PRIZES—		
	—	Scientific Activities Fund	100	
25	25	Grant to Field Studies Council	40	
				140
£11,244				£11,032

FUND

FOR THE YEAR ENDING 31 DECEMBER 1956

INCOME

1955					
£	£		£	£	£
		SUBSCRIPTIONS, ETC.			
	5,163	Subscriptions		5,216	
	509	Income Tax recovered from Covenanted Subscriptions		—	
	113	Entrance Fees		98	
	16	Donations		11	
	5,801			5,325	
	180	*Less* : Adjustments for Lapsed Membership (By-law 40)		167	
5,621					5,158
		PUBLICATIONS—			
	2,112	Sale of Publications – Current *Q.J.*		2,281	
	701	Sale of Publications – back numbers *Q.J.*		486	
	457	Reprints		351	
	317	Advertisements		514	
		Proceedings Toronto Meteorological Conference 1953—			
	289	Sales, etc.		45	
	3,876			3,677	
	500	Grant from Royal Society (excluding Library)		500	
4,376					4,177
		INVESTMENT INCOME—			
	260	Interest on Government Stocks (Gross)		260	
	236	Rent of Maisonettes	254		
	93	*Less* : Rates £83, Repairs *Nil*, Legal Expenses *Nil*	83		
	143			171	
403					431
		PHOTOGRAPHS—			
	16	Sale of Cloud Postcards		32	
	97	Clarke and Cave Copyright Fees	119		
	6	*Less* : Reproduction Costs	3		
	91			116	
107					148
		LIBRARY—			
	600	Grant from The Royal Society		400	
	1	Sale of old Library Books		50	
601					450
		AMOUNTS RECEIVED FROM OTHER FUNDS—			
	92	Life Composition Fund		200	
	15	British Rainfall Organization Fund (for Administration)		15	
107					215
	—	Grant from *Weather*			200
29		BALANCE, BEING EXCESS OF EXPENDITURE OVER INCOME—			253
£11,244					£11,032

RECEIPTS AND PAYMENTS OF SPECIAL FUNDS FOR 1956

	Balance at 31.12.55	Income for 1956	Expenditure for 1956	Balance at 31.12.56
	£	£	£	£
Awards Fund	10	—	—	10
British Rainfall Organization Fund	241	160	377 (a)	24
Life Composition Fund	428	1,087 (b)	1,010 (c)	505
Napier Shaw Memorial Fund	180	53	113	120
Royal Society Reprints Scheme	6	34	39	1
Scientific Activities Fund	110	173	148	135
Staff Pension Fund	513	514 (d)	486 (e)	541
Symons Memorial Fund	29	36	21	44
Hampton Brown Fund	—	1,000	—	1,000
	£1,517	£3,057	£2,194	£2,380

(a) Includes purchase of £300 3½% Treasury Stock 1979/81; (b) Includes £950 Defence Bonds Redeemed; (c) Includes purchase of £1,000 3½% Treasury Stock 1979/81 and transfer to General Fund £200; (d) Includes transfer from General Fund £387; (e) Includes purchase of £600 3½% Treasury Stock 1979/81.

Examined and found correct.

27 March 1957.

BALL, BAKER & Co.,
Chartered Accountants.

VALUATION OF INVESTMENTS AT 31 DECEMBER 1956

Market Value 31.12.55			*Market Value 31.12.56*	
£	£		£	£
		*General Fund**		
791		£1,400 2½% Consols	735	
1,056		£1,250 4% Consols x.d.	956	
689		£900 3½% Conversion Loan 1961 or after	635	
185		£200 4% Funding Stock 1960/90	175	
581		£649.1.7. Government of Commonwealth of Australia 4% Registered Stock 1955/70	535	
1,830		£2,000 4% Government of Commonwealth of Australia 4% Registered Stock 1966/68 x.d.	1,690	
5,132				4,726
		Scientific Activities Fund		
191		£250 3½% War Stock 1952 or after	176	
77		£100 3½% Conversion Loan 1961 or after	71	
875		£1,000 4% Trinidad Government Stock 1973/76	825	
1,143				1,072
		Staff Pensions Fund		
612		£800 3½% War Stock 1952 or after	564	
230		£300 3½% Conversion Loan 1961 or after	211	
650		£1,150 2½% Consolidated Stock	604	
761		£900 4% Consolidated Stock 1957 or after x.d.	689	
438		£500 4% Trinidad Government Stock 1973/76	413	
—		£600 3½% Treasury Stock 1979/81	465	
2,691				2,946
		Life Composition Fund		
452		£800 2½% Consolidated Stock	420	
612		£800 3½% Conversion Loan 1961 or after	564	
400		(£400 3% Defence Bonds)	—	
296		£350 4% Consolidated Stock 1957 or after x.d.	268	
550		(£550 3½% Defence Bonds Conversion Issue)	—	
—		£1,000 3½% Treasury Stock 1979/81	775	
2,310				2,027
		*British Rainfall Organization Fund***		
574		£750 3½% War Stock 1952 or after	529	
2,277		£3,097.10.0. British Transport 3% Guaranteed Stock 1978/88	2,122	
85		£100 4% Consolidated Stock 1957 or after	76	
—		£300 3½% Treasury Stock 1979/81	233	
2,936				2,960
		Napier Shaw Memorial Fund		
268		£350 3½% Conversion Loan 1961 or after	247	
338		£400 4% Consolidated Stock 1957 or after x.d.	306	
606				553
		Symons Memorial Fund		
497		£650 3½% Conversion Loan 1961 or after	458	
169		£200 4% Consolidated Stock 1957 or after	153	
666				611
£15,484				£14,895

* General Fund also includes the Freehold of 49 Cromwell Road, S.W.7.

** British Rainfall Organization Fund also includes the Freehold of 62 Camden Square, N.W.1.

Examined and securities seen.

27 March 1957.

BALL, BAKER & Co.,
Chartered Accountants.

Proceedings at meetings of the Society

Symons Memorial Lecture, 3 April 1957

Professor C. F. Powell, F.R.S., Melville Wills Professor of Physics, University of Bristol, delivered the Symons Memorial lecture, his subject being ' Experiments at high altitudes with free balloons.'

Annual General Meeting, 17 April 1957

Dr. R. C. Sutcliffe, O.B.E., F.R.S., President, in the Chair. The minutes of the Annual General Meeting held on 25 April 1956 were confirmed and signed.

In presenting the Report of the Council for the year 1956 the President referred to the main items of interest.

The financial position of the Society was giving rise to anxiety owing to the action of the Inland Revenue in not paying tax rebate on covenanted subscriptions. The last two years had shown a deficit.

The programme of meetings held by the Society had, as in former years, covered a wide field. The third two-day Summer Meeting held at Aberystwyth in honour of Sir David Brunt, Sec. R. S., whose seventieth birthday fell that year, had been very successful.

The inflow of communications for the *Quarterly Journal* had somewhat diminished, but the Society's publications had maintained their high standard and their thanks were due to the Editor, Editing Committee, Proof Readers and the *Weather* Committee.

The Membership of the Society was satisfactory and continued to increase, though at a somewhat slower rate. The Canadian branch and the centres in Scotland and Manchester continued to make good progress.

The President referred with appreciation to the services of the Staff during the year.

62 Camden Square, the property administered under the British Rainfall Organization Fund, was to be sold and, by arrangement with the St. Pancras Borough Council, the Observing Station transferred to the park in Camden Square.

The Treasurer reported on the financial section and drew attention to the bequest of £1,000 from the late Mr. A. Hampton Brown, formerly Assistant Secretary to the Society. The deficit of £253 shown on the year's working was entirely due to the non-payment of tax rebate on covenanted subscriptions as already referred to by the President.

The President moved that the Report of the Council be received and adopted and that it be printed in the *Quarterly Journal*. The motion was unanimously carried.

Prof. P. A. Sheppard proposed a vote of thanks to the President and Council for the year 1956. He expressed appreciation of the volume of work which had been accomplished and said that he felt possibly a very useful purpose would be served by reviving the ' Society News.' Mr. E. Gold seconded the vote of thanks.

The Treasurer rose to congratulate the President on his election into the Fellowship of the Royal Society. This was received with acclamation.

The Treasurer proposed the following annual motion on investments, submitted by the Council in accordance with By-law 86 :

> ' That a sum of up to £8,000 may be invested in non-Trustee stocks during the period of twelve months until the next Annual General Meeting'.

He explained that the amount which would be thus made available for investment was equivalent to just over half the present invested Capital of the Society, excluding the British Rainfall Organization Fund.

The motion was seconded by Mr. J. M. Craddock and carried *nem. con.*

The President then announced the following awards:

(a) *Symons Memorial Gold Medal for 1957* :

To Professor Erik Palmén, D.Ph., Institute of Meteorology, University of Helsinki, a member of the Academy of Finland, for his outstanding work in the field of synoptic aerology, as given in the following citation :

Professor Erik Palmén's scientific contributions over many years, especially in synoptic aerology, have been a model of critical analysis and lucid discussion. His more recent studies have been concerned with tropical hurricanes and the large-scale disturbances of the westerlies, the jet stream and frontal regions, the momentum and energy balance of the tropical atmosphere, and momentum exchanges at the earth's surface. He has provided convincing quantitative evidence of a direct meridional circulation in low latitudes and his well-known kinematical model of the general circulation of the atmosphere is a valuable and realistic synthesis of statistical and synoptic evidence.

His critical judgement and uncompromising thoroughness have established Professor Palmén as a leading world figure in contemporary meteorology.

The medal was then presented to His Excellency the Finnish Ambassador, Mr. Sakari Tuomioja, who received it on behalf of Prof. E. Palmén, and read the following message of thanks from him :

For the great honour conferred upon me by the Council of the Royal Meteorological Society through the award of the Symons Memorial Medal for 1957, I express my deep gratitude.

Feeling both pride and humility when accepting this high award, I am aware of my indebtedness to all those numerous friends and colleagues in different countries who have helped me and supported my works during almost thirty years of international collaboration. Through this continuous support it was possible for me to select problems of general importance for my investigations and to utilize the rapidly improving world-wide meteorological observations. Also, the great inspiration which I have received from many British scientists and from numerous papers published by the Royal Meteorological Society fills me with deep gratitude.

The extraordinary recognition shown me today by one of the world's oldest and foremost meteorological societies is always going to be a source of great encouragement during my further work in meteorology and adjacent fields.

(b) *Darton Prizes for 1956* :

Home Awards : (1) Mr. A. J. Drummond for his paper ' A contribution to absolute pyrheliometry ' (*Q.J.*, **82**, p. 481, 1956); (2) Dr. J. F. Nagel for his paper ' Fog precipitation on Table Mountain ' (*Q.J.*, **82**, p. 452, 1956).

Canadian Awards : (1) Mr. B. W. Boville and Mr. M. Kwizak for their paper ' 500-mb prediction by graphical techniques ' (being published in *Canad. Branch R. Met. Soc. Pap.*, **7**, No. 2); (2) Mr. G. A. McKay for his paper ' A method of minimum temperature prediction ' (being published in *Canad. Branch R. Met. Soc. Pap.*, **7**, No. 3).

Dr. R. C. Sutcliffe, O.B.E., F.R.S., then delivered the Presidential Address, his subject being ' An outlook on meteorology and the Society.' Dr. B. J. Mason proposed a vote of thanks to the President for his Address and asked that he would allow it to be printed in the *Quarterly Journal*. Mr. F. H. W. Green seconded the vote of thanks. The meeting was declared closed at 7 p.m.

Additional Meeting, 15 May 1957

A joint discussion meeting with the Royal Statistical Society on ' Analysis of geophysical time series ' was held at the London School of Hygiene and Tropical Medicine. Prof. E. S. Pearson, President of the Royal Statistical Society, was in the chair. There was an attendance of about 100 members of both Societies and guests.

The following papers were read and discussed :

' An analysis of the slower temperature variations at Kew Observatory by means of mutually exclusive band pass filters ' by J. M. Craddock, ' Notes on the specification of atmospheric turbulence ' by H. Charnock and ' Some applications of time series analysis to atmospheric turbulence and oceanography ' by S. Rushton and J. Neumann.

The above papers together with the contributions to discussion will be published in the *Journal of the Royal Statistical Society* (Series B).

Additional meeting, 23-24 May 1957

Joint symposia with the Royal Astronomical Society were held in their rooms, the topic being ' Planetary atmospheres and convection in rotating fluids.' About 110 members of both Societies and many visitors from overseas were present at the meeting which consisted of four symposia :

First symposium, Chairman : Sir Edward Bullard, F.R.S. Introduction by Sir Harold Jeffreys, F.R.S., outlining the history of problem; Prof. Z. Kopal on ' The basic observational data on planetary atmospheres '; Prof. P. A. Sheppard on ' The general circulation of the earth's atmosphere ' and Dr. H. U. Sverdrup on ' The circulation of the oceans.'

Second symposium, Chairman : Sir Harold Jeffreys, F.R.S. Dr. E. Lorenz on ' The circulation of the atmospheres of the outer planets '; Dr. G. B. Tucker on ' Transfer processes in planetary atmospheres '; Dr. R. Hide on ' Experiments on convection in rotating liquids '; Prof. M. J. Lighthill on ' The general theoretical aspects of the problem as raised in the preceding papers '; discussion.

Third symposium, Chairman : Prof. M. J. Lighthill, F.R.S. Dr. R. M. Goody on ' The thermal structure of Mars '; Prof. H. C. Urey on ' The escape of planetary atmospheres '; Dr. R. C. Sutcliffe on ' Influence of condensation on the circulation of planetary atmospheres '; Mr. J. S. Sawyer on ' Jet streams as a feature of the earth's atmosphere '; discussion.

Fourth symposium, Chairman : Prof. P. A. Sheppard. Mr. T. V. Davies on ' The theory of the experiments on convection in rotating liquids '; Prof. J. Charney on ' Theory of the general circulation of the atmosphere '; Dr. E. T. Eady on ' The physical parameters characterizing the dynamics of planetary atmospheres '; discussion, followed by a summing up by Prof. P. A. Sheppard.

The meeting was organized on behalf of the Society by a working party consisting of Mr. T. V. Davies (co-ordinator), Dr. R. M. Goody and Prof. P. A. Sheppard, to whom Council wish to record their thanks. A fuller account of the meeting has appeared in *Weather* and a report will be published in *Nature*.

Ordinary Meeting, 29 May 1957

The following candidates were declared to have been elected on 17 April by the Council :

FELLOWS

BENNETT, Kenneth Graham, Northern A.I.S. Centre, Manchester Airport, Wythenshawe, Manchester, Lancs.

FEAVER, Russell John, B.A., 75 Sandlands Road, Walton-on-the-Hill, Tadworth, Surrey (*transfer from Student Membership*).

FORTESCUE, Mrs. Gillian Grace, B.Sc., 28 Rectory Close, Byfleet, Weybridge, Surrey (*transfer from Student Membership as Miss G. Strange*).

HODGKISS, Peter Keith, Coton House, Churchover, Nr. Rugby, Warwickshire (*transfer from Student Membership*).

MATTOCK, John Edward, 21 Kensington, Bath, Somerset.

ROGERS, David Aveline, 39 Courthorpe Road, London, N.W.3.

STODHART, Arthur Henry, A.M.I.E.E., 80 Mount Pleasant, Aspley Guise, Bletchley, Bucks.

TARKOWSKI, Capt. Czeslaw, c/o West African Aircraft Corporation, Lagos Airport, Lagos, Nigeria.

WILLSON, Ian Reginald, 21 Cavendish Heights, Cavendish Road, Yeoville, Johannesburg, S. Africa.

Student Members

Hall, Barry Peter Noel, 9 Jackson Road, Clacton-on-Sea, Essex.
Patanasiri, Aroon, Meteorological Office, Royal Thai Air Force, Donmaung, Bangkok, Thailand.

The following candidates were then declared elected on 15 May by the Council :

Fellows

Ball, Peter Edward John, ' Cottesmore,' King's Way, Harrow, Middlesex (*transfer from Student Membership*).
Davies-Scourfield, Margaret Gwynne, F.R.A.S., Annington House, Steyning, Sussex.
Houghton, John Theodore, M.A., D.Phil., Ambarrow Court, Sandhurst, Camberley, Surrey (*transfer from Student Membership*).
Ledbury, Adrian Robin, 21 Bathwick Hill, Bath, Somerset.
Nicholson, Willian Joseph, North Lodge, Moor Park, Preston, Lancs.
Parker, Albert Edward, B.Sc., A.M.Tech. I.,49 Poynters Road, Dunstable, Beds.
Probert-Jones, John Richard, B.A., 26 St. Michael's Road, Bedford (*transfer from Student Membership*).
Vineall, Alec Frederick, Norton Priory Lodge, Church Norton, Selsey, Chichester, Sussex.
Wolf, Peter Otto, B.Sc., A.M.I.C.E., A.M.I.W.E., 13 Burgh Heath Road, Epsom, Surrey.

Foreign Members

Estoque, Mariano A., Ph.D., Arctic Meteorology Research Group, McGill University, Montreal 2, Canada.
Sekihara, Kyo, D.Sc., Aerological Laboratory, Meteorological Research Institute, Mabashi, Suginami-ku, Tokyo, Japan.

Student Member

Qureshi, Mohammed Yousuf, 243 Cromwell Road, London, S.W.5.

Canadian Branch

Fellows

Basset, Gordon, L., M.A., 281 Humbercrest Boulevard, Toronto, Canada (*transfer from Associate*).
Finlayson, Ronald Campbell, 2996 West 20th Avenue, Vancouver 8, B.C., Canada.
Held, Isser Edward ,B.Sc., 315 Bloor Street West, Toronto 5, Ontario, Canada.

Associates

Boucaud, Albert Andrew, M.A., 315 Bloor Street West, Toronto 5, Ontario, Canada.
Dexter, Eric Henry Vincent, B.Sc., 315 Bloor Street West, Toronto 5, Ontario, Canada.
McCabe, Henry John, B.A., 315 Bloor Street West, Toronto 5, Ontario, Canada.
Macfarlane, Mona Aylmer, B.Sc., Department of Geography, McGill University, 539 Pine Avenue W, Montreal, P.Q., Canada.
Robert, André J. M. A., 315 Bloor Street West, Toronto 5, Ontario, Canada.
Shewchuk, Orest, B.Sc., 315 Bloor Street West, Toronto 5, Ontario, Canada.
Watson, Warren Blake, B.Sc., 315 Bloor Street West, Toronto 5, Ontario, Canada.

The meeting took the form of a joint discussion with the National Institute of Oceanography on ' Long waves and storm surges ' and consisted of two sessions with the President, Dr. R. C. Sutcliffe, F.R.S., in the chair. The afternoon session was opened by Mr. F. R. Snodgrass (Scripps Institution of Oceanography) on ' The measurement of long waves on the ocean ' followed by Prof. W. H. Munk (Scripps Institution of Oceanography) on ' The long-wave spectrum ' and Dr. H. U. Roll (Deutscher Wetterdienst, Seewetteramt) who spoke on ' Forecasting surface gravity waves ' and illustrated his talk with a film.

The evening session was opened by Dr. D. Lee Harris (United States Weather Bureau) who showed a film and spoke on ' Hurricane surges ' followed by Prof. P. Groen (Royal Netherlands Meteorological Institute) on ' North sea surges ' and Dr. P. Welander (Meteorological Institute, Stockholm) on ' A programme for the numerical prediction of storm surges '. Each contribution was followed by discussion. A fuller report of the meeting has been published in *Weather*.

Reviews

Allgemeine Meereskunde : Eine Einführung in die Ozeanographie. Von G. Dietrich mit Beiträgen von K. Kalle. Berlin (Gebr. Borntraeger), 1957. Pp. viii, 492; 223 Figs.; 16 Plates; 7 Charts in pocket; 63 Tables. 56 DM.

It is now 46 years since Otto Krummel's *Handbuch der Ozeanographie* appeared, and already *The Oceans* by Sverdrup, Johnson and Fleming is 15 years old. The former is still much consulted, and the latter is of course an invaluable compendium; but the development of oceanography has been so great in recent years as to call for a new book which should present *inter alia*, between two covers, some account of the modern advances in knowledge of waves, in instrumental techniques, and in other departments of marine physics which have been the subject matter of a host of papers published during the last 15 years in the pages of many journals.

The elegant book before us is from the pen of a man whose contributions to oceanography, ranging over a surprisingly wide field, have earned the greater esteem because of his sure practical touch, where oceanography serves fisheries and navigation for instance. The reviewer has heard regret expressed that the present work does not make such a worldwide presentation of regional oceanographical information as either of its two illustrious predecessors named above, but such a criticism is hardly a fair one. Practitioners of the science are much more numerous nowadays; the ambit has expanded amain, and the tempo of advance has quickened so enormously, thanks in large part to modern methods of observation and data analysis subserved by electronics, that a really good book (and this is such !) merits a very warm welcome indeed.

Nothing of comparable scope to this present book – unless perhaps in the Russian language – has appeared since the time *The Oceans* was published, and the reader who may already be familiar with Kalle's *Der Stoffhaushalt des Meeres*, will be pleased that its distinguished author has contributed so much. It can be revealed as a great tribute to the author's industry that, when an intending co-author was unable to undertake the treatment of tides for inclusion in the book, Dietrich shouldered the extra work himself.

It is an inevitable and unfortunate feature of such a book as this, that a mere recital of contents supplemented by a brief description of some of the illustrations, would absorb as much space as is usually allotted for a review. We shall give no such recital, impressive though that would be. In compass and get-up the book is astonishing, and a mere turning of the pages from end to end leaves one in no doubt that the author's aim expressed in his foreword has been realized – to give his readers some insight into all branches of oceanography. To have entertained such an aim was indeed ambitious; to have realized it is an achievement which not only required great industry but attests a truly remarkable acquaintance with the discovered facts and literature of oceanography. Only so could such a connected story as is told to the reader of these pages have been possible.

Without going into details it is to be remarked that the book begins with a geomorphology of the ocean which deals with the topography of its floor, its sediments, and the origin of its bottom features. There follows a treatment of the physical and chemical properties of the waters of the seas – characteristics which are so important in respect of biological production and of fisheries. A following portion of the book is devoted to a description of oceanographical instruments and methods of measurement. Then comes the treatment of important individual topics with attention given to the heat-budget of the oceans; the distribution of temperature, salinity, density and ice over the oceans; their "Stoffhaushalt"; the theories of ocean currents; surface waves; internal waves; long waves and tidal waves. The last chapter treats regional oceanography, goes into the very important and topical subject of the circulation in the ocean depths, and discusses the hydrographical regions of the world oceans. At the end of the book is an excellent 14-page literature list classified under 10 main topics, a source-index detailing whence the diagrams were taken, a name index, and a subject index.

The 7 excellent folded charts in the back-cover pocket present respectively :— the main depth divisions of the oceans; the distribution of sediment types; the surface temperature of the oceans in anomaly form; the surface salinity distribution likewise presented; the surface currents of the oceans; the distribution of half-daily and other tides over the oceans, and finally, the distribution and maximum range of the diurnal tides.

J.N.C.

The structure of turbulent shear flow. By A. A. Townsend. Cambridge University Press, 1956. Pp. xii, 315; 103 Figs.; Tables. 40s.

For the first time, much of the work on the nature of the turbulence in shear flow carried out in the last decade or so is brought together in this monograph. And surely, few can be as qualified for the task as the author who, with his co-workers at the Cavendish Laboratory, has done so much of this pioneer work. Practically all the experimental work on shear flows discussed in this book has been carried out in the controlled conditions of wind tunnels. Consequently many of these experimental results will not be of immediate benefit to the meteorologist, but indirectly those theories which are shown to be supported by wind-tunnel results must be of great help to the theoretician and in turn to the experimentalists in our field. In particular, several important concepts and theories are considered and put to the test; these may well form the main foundations in future theories of atmospheric turbulence near the ground. They are the principles of similarity of flow at high Reynolds numbers, the analysis of self-preserving flows and the equilibrium hypothesis for large eddies.

The introduction describes the effects of anisotropy on the correlation functions and energy spectra. Since this is done in terms of simple eddy structures, it should provide a useful guide to the descriptive analysis of results in more complex circumstances. Chapter 2 gives the equations of motion in shear flow; chapter 3 discusses the return to isotropy from an anisotropic state. In analogy with the cascade process of homogeneous isotropic turbulence, it is deduced that the smallest eddies in shear flow are isotropic. The natures of wake flow, of rapidly distorted homogeneous fields and of unrestricted shear flows are analysed in the light of the theories mentioned above. A good number of these results will be of interest to meteorologists. However the same cannot be said of four of the twelve chapters, namely 7, 8, 9 and 12, except in so far as they add evidence in support or opposition to general theories. They are concerned with flow in pipes, flow round cylinders and other non-meteorological type flows.

Chapters 10 and 11 deal with boundary-layer flows and as such are of meteorological importance. The method of attack and the deductions about the fluxes of momentum and energy, the conditions and development of self-preserving flows are given in some detail.

This brief summary of some of the topics treated will indicate that this book should provide a very useful source of ideas to those tackling this difficult problem.

F.B.S.

Hygrometry. By H. Spencer Gregory and E. Rourke. London (Crosby Lockwood), 1957. Pp. xv, 254; 119 Figs.; 17 Tables. 36s.

Although this book appears to attempt to satisfy the need for an advanced and scholarly treatise on hygrometry, it does not do so. Even though it has some very good points, the reviewer regretfully feels it is better to state simply that the book is not generally recommended, rather than to detail its defects.

Rather unusually, the preface is a historical survey of the subject, and the introduction is an account of the main features of the useful hygrometers. Succeeding chapters then give a theoretical background. A specialist who wishes to read details of the latest hygrometers, will, if he is a competent physicist able to read the book in a fully critical way, find it useful and stimulating.

In its stimulating nature the book reflects the character of Dr. Spencer Gregory whose work on the problems of hygrometry has been very extensive, and whose recent death is recorded with great regret.

A.W.B.

The ways of the weather. By J. S. Sawyer. London (Adam and Charles Black), 1957. Pp. vi, 97; 46 Figs.; 18 Plates. 12s. 6d.

Not the least of the problems facing the ' Weather Man ' is the explanation of the mysteries of his craft to an interested and frequently uninformed public. Recent advances have revealed such complexities in the atmosphere that the writing of even elementary textbooks on the subject is best left to the professional meteorologists. He who lives with the atmosphere from day to day is less likely to make those mis-statements which, once in print, are often quoted over and over again to the general detriment of the science. Nor is he likely to under-rate the difficulties which face those who may wish later to delve more deeply into the subject.

Mr. Sawyer's book *The ways of the weather* should therefore appeal to a wide public. It is written for the beginner and presumes no knowledge of any branch of science. Like other writers, the author seems to have found some explanations difficult in the absence of mathematics and, on such occasions, to have oversimplified his theme, e.g. in the section on the geostrophic wind relation. Nevertheless he covers a wide field in the 95 pages of text and manages to sustain our interest throughout. He writes with an engaging frankness, particularly in the section entitled ' The future of forecasting.

The diagrams are excellent and well-selected. They assist the text adequately without losing their clarity in a fog of irrelevant detail. There are some excellent photographs too, though the book gets off to an unimaginative start with a wretched picture of an apparently formless balloon being released from the deck of a weather ship in dock ! And surely there is a need for a selected reading list in a book which is certain to whet many a beginner's appetite for more. Can we hope for such an improvement in later editions of this excellent little work ?

J.B.R.

Advances in geophysics, Vol. 3. Edited by H. E. Landsberg. New York (Academic Press), 1956. Pp. x, 378; Figs. £3 10s. 6d.

The editorial committee of this series has the ambition to make ' The broad appeal and promise of joint discussion, regardless of specialization, covering the whole field of Geophysics . . . . .' This is a wholly admirable ambition and the committee has once again managed to achieve it. The result is a well-produced and satisfying set of reviews which should give pleasure to any geophysicist.

This gives the reviewer very little scope, for it is unrewarding to look for minor blemishes in carefully-produced reviews by recognized experts. The titles and authors should convey the quality of the book, so they deserve to be listed :

' Arctic ice island research ' by A. P. Crary; ' Recent developments in the study of the polarisation of sky light ' by Z. Sekera; ' Subcontinental structure in the light of seismological evidence ' by P. Byerly; ' Heatflow through the Deep Sea Floor ' by E. C. Bullard, A. E. Maxwell and R. Revelle; ' The interior of the Earth ' by J. A. Jacobs; ' Subsurface geophysical methods in ground-water hydrology ' by P. H. Jones and H. E. Skititzke; ' Geophysical research with artificial Earth satellites ' by S. G. Singer.

The article by Bullard, Maxwell and Revelle is entertaining. The geophysicist can boast a real advantage over his laboratory-bound colleagues when he can get useful results by projecting 15 ft darts into the ocean bed at depths up to 5 km. He cannot, of course, be too sure what has happened to them but that is part of the attraction of the game.

To the reviewer the most valuable contribution was that by Jacobs. The reader who has been baffled and perhaps intimidated by bold assertions as to what goes on in the interior of the earth, will be relieved to find reference to a *Glossary* by Birch, a few examples of which are (accepted meaning in parenthesis) : certain (dubious); positive proof (vague suggestion); pure iron (uncertain mixture of all the elements).

Singer's article is, as the author points out, mainly for propaganda purposes. Unfortunately it is at times too superficial to convince a doubtful customer for artificial satellites and a more thorough analysis of less material would probably have been more effective. For example, the Rossby and Palmén circulation schemes, polar-front cyclones, jet-stream meanderings and model experiments jostle together in eight lines, but the help which the satellite will give to these exceptionally difficult problems does not relate directly to any of the questions which dynamical meteorologists are currently asking.

R.M.G.

Obituary Notices

Dr. GUSTAV SWOBODA

To many a scientist the day comes when he must make a choice – whether to continue in a job where he can carry on his own original scientific research or to accept a new post in which the administrative duties will absorb all his energies. This must have been the dilemma which faced Dr. Gustav Swoboda when in 1938, after a career of research, teaching and practical meteorology, he was offered the post of Secretary-General of the International Meteorological Organization. Only a person who has served in an international organization can appreciate how time-absorbing, albeit fascinating and often exciting, the work can be. Little wonder then that after 1938 Dr. Swoboda's direct contributions to meteorological research were not of great significance. Many will consider, however, that this loss was more than compensated by his indirect contributions to the science of meteorology as the executive head of the IMO and of its successor the World Meteorological Organization, for research workers in meteorology are probably more dependent than any other scientists on an efficient international organization to ensure a reasonable uniformity of method in the observations which provide the raw material for their work.

Dr. Swoboda was born on 7 September 1893, and after studying meteorology, physics and philosophy at the King Charles University in Prague received the degree of Doctor of Philosophy in 1920. This was an exciting period for meteorologists and Dr. Swoboda was soon in the forefront of those who pursued the new ideas on frontal theory of the Bergen school. In 1924 he published an important work of 178 pages, in collaboration with Professor Tor Bergeron, on waves and vortices on a quasi-stationary surface. He was deeply interested in the practical application of the new theories and wrote several papers and books in Czech on weather forecasting, and especially on its development to meet the needs of aviation. His last major contribution to the scientific literature was in 1938 when he completed a new version of Chromov's *Introduction to synoptic weather analysis* in collaboration with Dr. M. N. Koncek. This was no mere translation from Russian to German (although that in itself would have been a considerable undertaking) but an adaptation of the work to the problems of Central Europe.

In his subsequent international work, Dr. Swoboda became one of the best known and best loved meteorologists in the world. His unassuming presence at countless international gatherings was an assurance of their success and his unusual linguistic accomplishments (he was bilingual in Czech and German, spoke French and English fluently, and had a useful knowledge of several other European languages including Dutch) encouraged meteorologists of many nationalities to turn to him as counsellor and friend.

During the second World War he succeeded in holding the IMO together and shortly after its termination he was involved in the complicated negotiations which resulted in the birth of WMO – a more powerful inter-governmental organization and a Specialized Agency of the United Nations. From then until his retirement in 1955, Dr. Swoboda worked endless hours building up the WMO Secretariat and setting in motion the various WMO constituent bodies – the Regional Associations and the Technical Commissions. By that time he was a tired man, but after a year as Professor of Meteorology at the University of Istanbul he became completely rejuvenated. When he visited the WMO Secretariat on leave in July 1956 he gave an impression of great vitality and spoke of his plans for starting some original research during the following scholastic year. These plans were not to materialize, for on 4 September 1956 he passed away, a few days after an operation. He left behind a widow and one son – and countless friends all over the world.

Dr. Swoboda was elected a Foreign Member of the Royal Meteorological Society on 17 June 1953.

O.M.A.

Mr. A. R. SHAW

Mr. Alan Roger Shaw died at Edmonton, Canada, on 8 May 1956, after a brief illness, at the age of 39. Born in Halifax, Nova Scotia, he spent his early youth at Berwick, near Halifax. He joined the Royal Canadian Air Force in April 1941 and subsequently saw four years of active service in England. On his discharge from the R.C.A.F. in 1945 he joined the Meteorological Service of Canada as an observer, in which capacity he rendered valuable service at many stations in western Canada until his death. Mr. Shaw had an enthusiastic interest in meteorology which was manifested by his diligent and selfless devotion to duty, at times under conditions of physical hardship. These qualities earned for him a high reputation. He was elected an Associate of this Society on 21 February 1951.

R.L.

Dr. F. A. HENSON, B.Sc., Ph.D., F.G.S.

Dr. Frederick Arthur Henson, lecturer in the Department of Geology, University of Nottingham, died suddenly on 14 January 1957. Born at Wolverton, Bucks, on 30 June 1919, he entered the University of Reading in 1937. From 1940 to 1945 he was on active service, first as a fighter pilot in the Fleet Air Arm and later as Staff Meteorological Officer on the personal staffs of Admiral Somerville and Admiral Mountbatten in South-East Asia Command. Resuming his studies after distinguished war-time services, he completed his

degree and proceeded to research work in the University of Reading on the geology of the Channel Islands, for which he was awarded the Ph.D. degree of his University. In 1947 he was appointed lecturer in the University of Nottingham and developed courses in petrology and economic geology. His work in the Channel Islands continued and a series of important papers, some of which are still in manuscript, have emerged on the granites and the migration of trace elements associated with granitisation in Jersey. For the past three years Dr. Henson stimulated researches on the basalts and associated rocks of Iceland and with immense energy and enthusiasm he made two expeditions to study these rocks. At the peak of his career it seems tragic that the results of this important research had just reached the stage of publication. His writings covered a wide variety of topics and he was responsible for the teaching and development of the only complete course of study in Photogrammetry at present existing in British Universities. He was an active member of the Board of the Faculty of Pure Science, the University representative on the Outward Bound Trust and a Lieutenant-Commander in the R.N.V.R. His stimulus and influence on all sections of University life will be a great loss and his ever ready help to expeditionary projects will be sadly missed by the undergraduates of the Universites of Nottingham and Reading.

I.R.B.

Corrigendum to October 1954 issue

K. L. S. Gunn and T. W. R. East, p. 539—

In the last four lines of Table 4, giving back-scatter in water cloud, all the coefficients are too large by a factor of one million. E.g., the coefficient for $\lambda = 3{\cdot}2$ cm at 0°C should read $7{\cdot}93 \times 10^{-8}$.

The entries for snow in Tables 4 and 5 are based on the relationship $Z = 200\,R^{1{\cdot}6}$. More recent measurements of snowflake-size distributions indicate a relationship $Z = 2{,}000\,R^{2{\cdot}0}$. To take account of this change, (and to take in the corrigendum of October 1955, p. 653) all coefficients of $R^{1{\cdot}6}$ in snow entries in both tables should be increased by the factor 10·6 and the index changed from 1·6 to 2·0. E.g., the attenuation due to snow at $\lambda = 3{\cdot}2$ cm and −10°C should read $34{\cdot}8 \times 10^{-5}\,R^{2{\cdot}0} + 22{\cdot}9 \times 10^{-5}\,R$.

Tables 4 and 5 have been revised and brought up to date : mimeographed copies are available on application to the authors, Physics Department, McGill University, Montreal, Canada.

FOR PUBLICATION SEPTEMBER

Prepared under the auspices of the Royal Meteorological Society

CLOUD STUDY

A PICTORIAL GUIDE

By F. H. Ludlam and R. S. Scorer

D.I.C., F.R.Met.S. M.A., Ph.D., F.R.Met.S.

Foreword by R. C. SUTCLIFFE, O.B.E., Ph.D., F.R.S.

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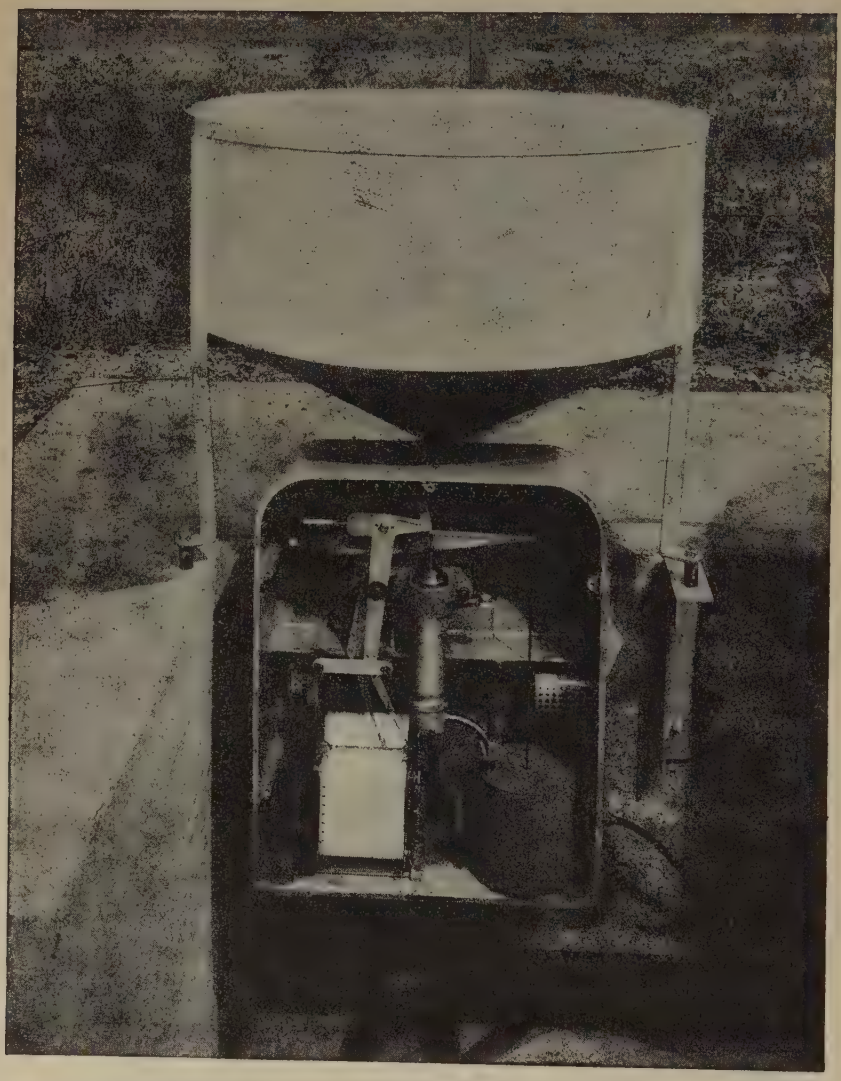

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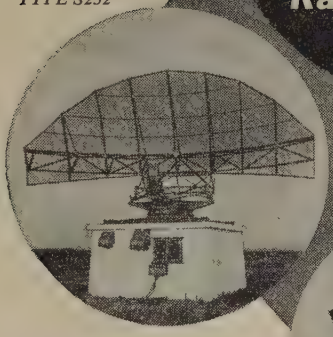

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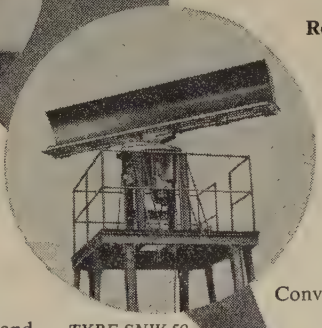

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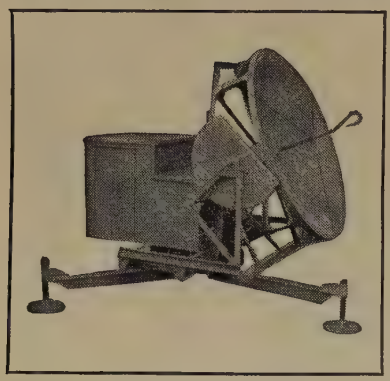

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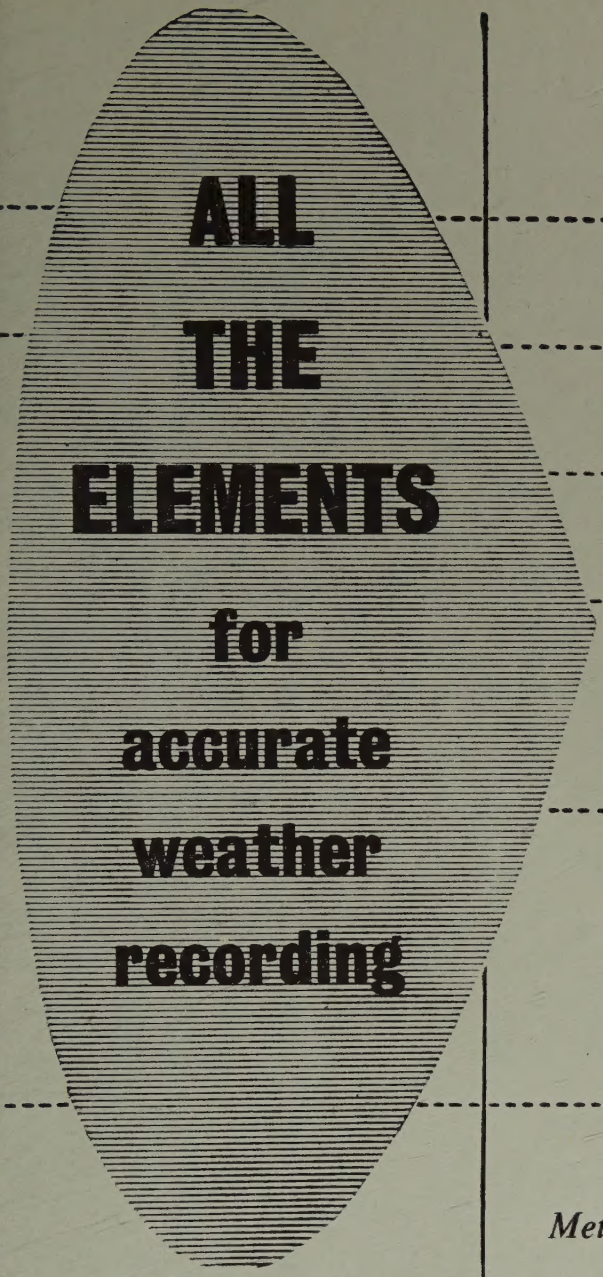

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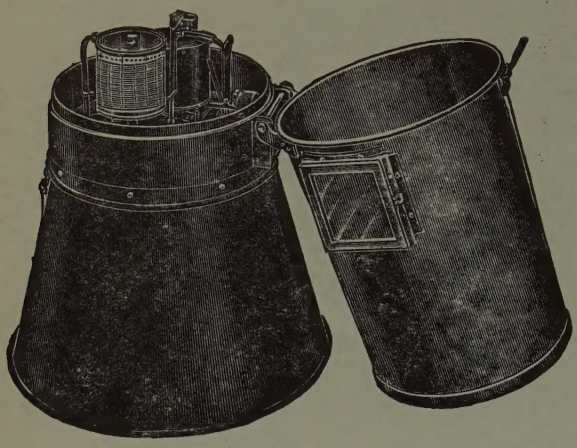

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